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To Our Readers

FOR fifty-five years the *Journal of Geology* has been under the inspiring and devoted editorships of Thomas Chrowder Chamberlin and Rollin T. Chamberlin. Under their guidance it has become an effective instrument for the dissemination of the results of geological investigations, presenting researches of fundamental importance. Now the editorship passes to Dr. F. J. Pettijohn, of the Department of Geology at the University of Chicago. He will have the active support of a group of leaders in the various fields of geology both in the United States and abroad.

The *Journal* will publish papers reporting fundamental research in all branches of geology. Contributions from any part of the world will be considered and may be written in English, or in French or German with an English summary. To facilitate publication, authors are urged to observe the requirements detailed on the next page.

Among the articles to appear in early numbers of the *Journal of Geology* are the following:

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"A Preliminary Report on Vertebrates from the Permian Vale Formation of Texas." By EVERETT C. OLSON

"Cave-in Lakes in the Nabesna, Chisana, and Tanana River Valleys, Eastern Alaska." By ROBERT E. WALLACE

"An Unusual Volcanic Dike and Grooved Lava at Auckland, New Zealand." By J. A. BARTRUM and E. J. SEARLE

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——— (1884*b*) Guidebook for excursions in the Big Ben Mountains, pp. 32, 33, New York, Macmillan Co.
SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.
———, and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76, 3 pls.
———, ——— (1929*b*) Revision of Ordovician stratigraphy of the Black Hills: Jour. Geology, vol. 37, pp. 680–695.

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EARLY TERTIARY FANGLOMERATE, BIG HORN MOUNTAINS, WYOMING

ROBERT P. SHARP

California Institute of Technology, Pasadena

ABSTRACT

Great accumulations of coarse bouldery gravel along the east base of the central Big Horn Mountains are composed almost wholly of debris derived from the pre-Cambrian core of the range. The name "Moncrief gravel" is proposed for this deposit. It has previously been described as Pleistocene glacial material, late Tertiary to earliest Quaternary bench gravels, or the coarse phase of an early Tertiary basin fill. The gravel was found to be gradational into fine-grained early Tertiary beds and to be separated from the pre-Tertiary rocks of the mountains by thrust faults. For these reasons the Moncrief gravel is identified as an early Tertiary, probably Eocene, fan deposit, formed as the Big Horn Mountains were progressively uplifted and thrust eastward during the Laramide Revolution. Early Tertiary glaciation may have played a part, but this is largely speculation.

INTRODUCTION

GENERAL STATEMENT

Thick deposits of coarse bouldery gravel with interbedded silt, sand, and arkose layers compose large piedmont ridges and spurs along the east front of the Big Horn Mountains. This gravel consists predominantly of granitic rock fragments derived from the pre-Cambrian core of the range and has its best development in Bald, North, and Moncrief ridges (fig. 3), which extend 2–3 miles east from the mountains and rise 1,500 feet above the general piedmont level.

Max Demorest (1938, p. 18) referred to these deposits as the "Clear Creek gravels," but Clear Creek is preoccupied several times over (Wilmarth, 1938, pp. 457–459), so the name "Moncrief gravel," from excellent exposures on Moncrief Ridge (fig. 3) in Sheridan County, is substituted. The best exposures are in high north-facing cliffs at the top of the ridge and on the slopes below; hence the north face of Moncrief Ridge (secs. 34 and 35, T. 54 N., R. 84 W., Sheridan quadrangle) is designated as the type locality. The present topographic surface determines the top of the unit, and its basal and lateral margins are gradational into finer-grained "Wasatch" beds except where the gravel lies with angular unconformity on the Eocene Kingsbury conglomerate and older formations or is in fault contact with pre-Tertiary rocks of the mountains. More detailed work may subsequently provide a better basis of delineation, but for the present the boundaries of the Moncrief may be determined by the lowermost and outermost gravel bed, consisting primarily of pre-Cambrian rock debris. The age

is early Tertiary, with Eocene a strong probability; and the maximum thickness observed on Moncrief Ridge is 1,400 feet.

Brief investigation of these deposits was made in 1940 and 1946, incidental to study of erosion surfaces along the east flank of the Big Horn Mountains. Field time totals 7 weeks.

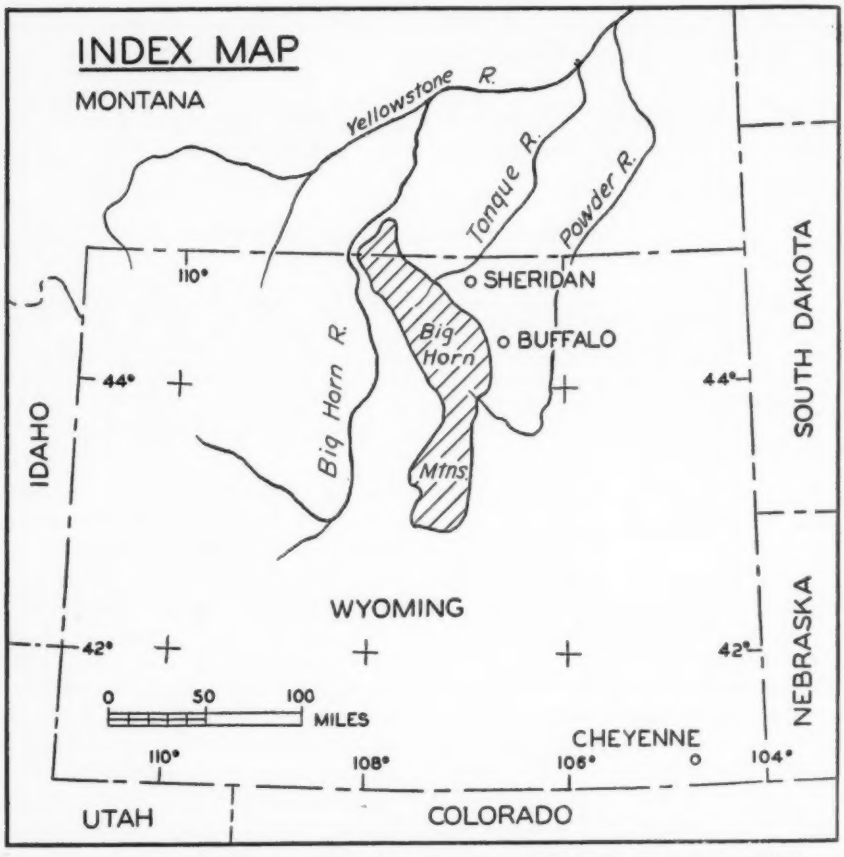


FIG. 1.—Area described lies at east base of Big Horn Mountains between Sheridan and Buffalo

GEOLOGICAL SETTING

The Big Horn Mountains are the front range of the Middle Rockies in north-central Wyoming (fig. 1). They consist of a major anticlinal uplift trending northwest, with an exposed pre-Cambrian core flanked by upturned Paleozoic and Mesozoic strata. Laramide thrusting is directed eastward in the central section of the range, westward in the north and south sections (Demorest, 1941, pp. 165–166; Chamberlin, 1940, p. 680), and southward at the southern end (Love, 1940, p. 1934; Tourtelot, 1946). In the adjoining basins Mesozoic and Paleozoic rocks are largely buried beneath early Tertiary continental deposits. Topo-

graphically the central part of the range consists of a backbone of high peaks (Cloud Peak, 13,165 feet) flanked by a broad rolling subsummit upland at 7,500–9,000 feet (fig. 2). Imposing parapets developed by erosion of sharply upturned Paleozoic beds form the east front of the range and separate the subsummit upland from the piedmont slope, which lies at 4,500–6,000 feet and truncates tilted Mesozoic and more nearly horizontal Tertiary strata. The backbone of the range has been heavily scoured by late Pleistocene glaciers, which advanced onto the subsummit upland but failed to reach the east base of the range (Darton, 1906*a*, pl. 26).

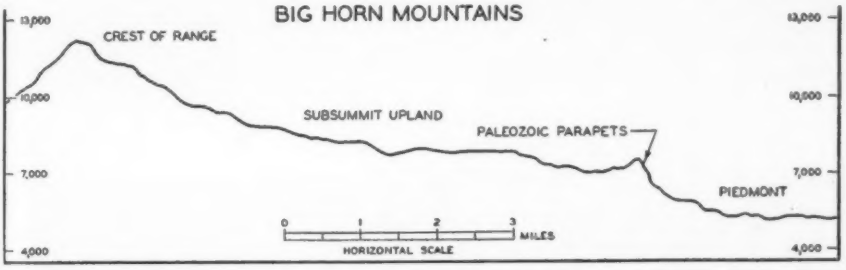


FIG. 2.—Topographic profile, east side of Big Horn Mountains

REVIEW OF PREVIOUS WORK

R. D. Salisbury and Eliot Blackwelder (1903, pp. 221–223) first described the bouldery accumulations at Bald and North ridges and tentatively suggested that they might be glacial deposits of earlier date than the late Pleistocene glaciation recognized higher on the mountains. Blackwelder (1915, pp. 313–315) later compared the deposits of Bald Ridge with coarse fan or outwash gravel on the Black Rock erosion surface (early Pleistocene) in western Wyoming. N. H. Darton (1906*a*, pp. 68–70, pl. 47; 1906*b*, pp. 8–9) mapped all occurrences of the Moncrief gravel and concluded that they were remnants of a late Tertiary or earliest Quaternary bench gravel which at one time extended all along the east-central front of the range. W. C. Alden (1932, pp. 41–44) gives the most thorough treatment of the deposits composing Bald, North, and Moncrief ridges and suggests that they are remnants of great lobate terminal moraines built by early Pleistocene glaciers which flowed from mountains onto the surface of the Flaxville Plain (No. 1 bench). The late Max Demorest (1938, pp. 22–23) suggested two possible interpretations for the Moncrief gravel: (1) It may be a coarse, mountain facies of the widespread Tertiary deposits which filled the intermontane basins and built up the Great Plains, or (2) it is the remnant of local fans deposited in response to local and temporary conditions after the plains surface was largely developed, perhaps in the early Pleistocene. As a result of later field work, he[1] dismissed the second possibility and adopted the view that the Moncrief gravel was gradational into the underlying Eocene Kingsbury conglomerate and also continuous in its uppermost part with Tertiary(?) gravel on the subsummit upland (fig. 2) of the Big Horn Mountains. His published report

[1] Personal letters (1940).

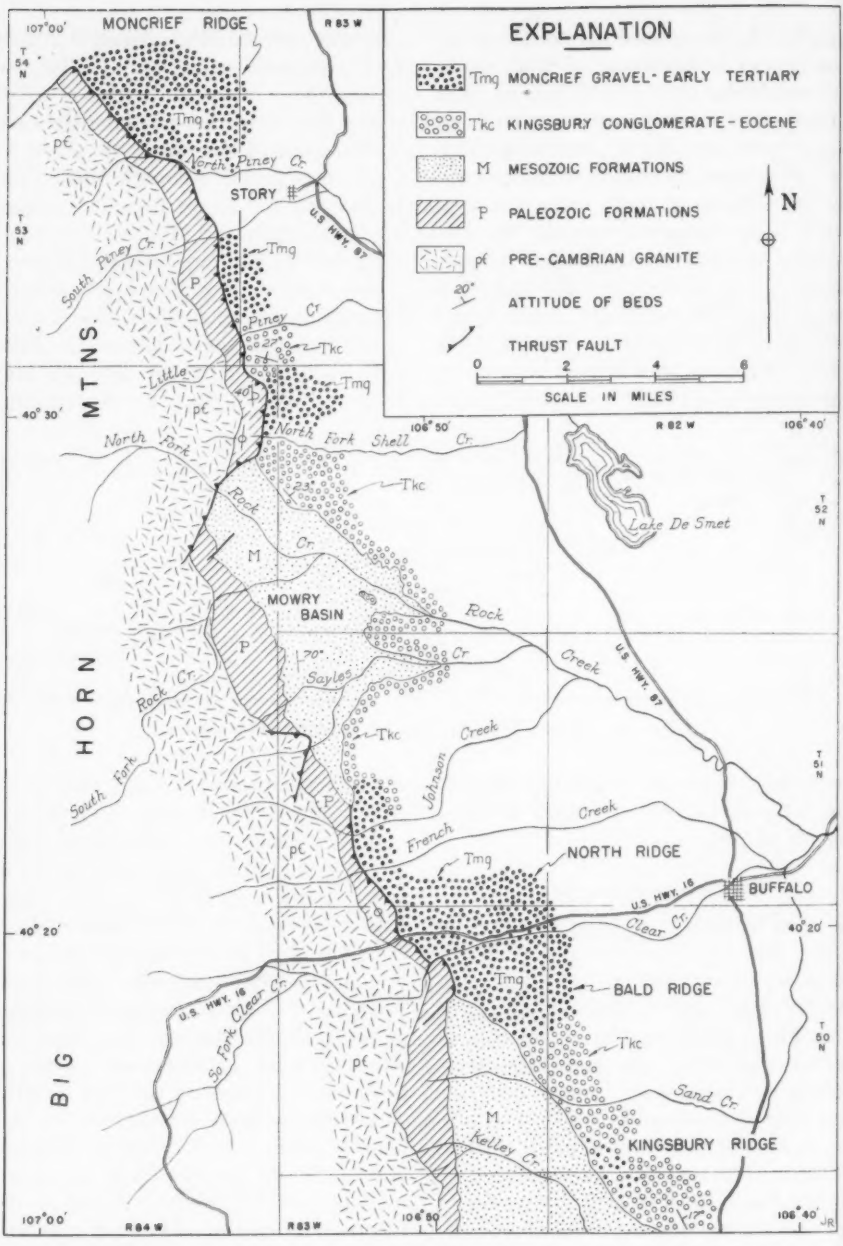


FIG. 3.—Geological map along east base of central Big Horn Mountains, modified from Darton and Demorest.

PLATE I

A

B

A, Boulder beds at top of Moncrief gravel in type section on north side of Moncrief Ridge. Arkose, sandstone, and siltstone in middle of exposure. About 200 feet of beds shown.

B, Large boulders of pre-Cambrian granite and gneiss in Moncrief gravel north of Little Piney Creek. View northeastward.

PLATE II

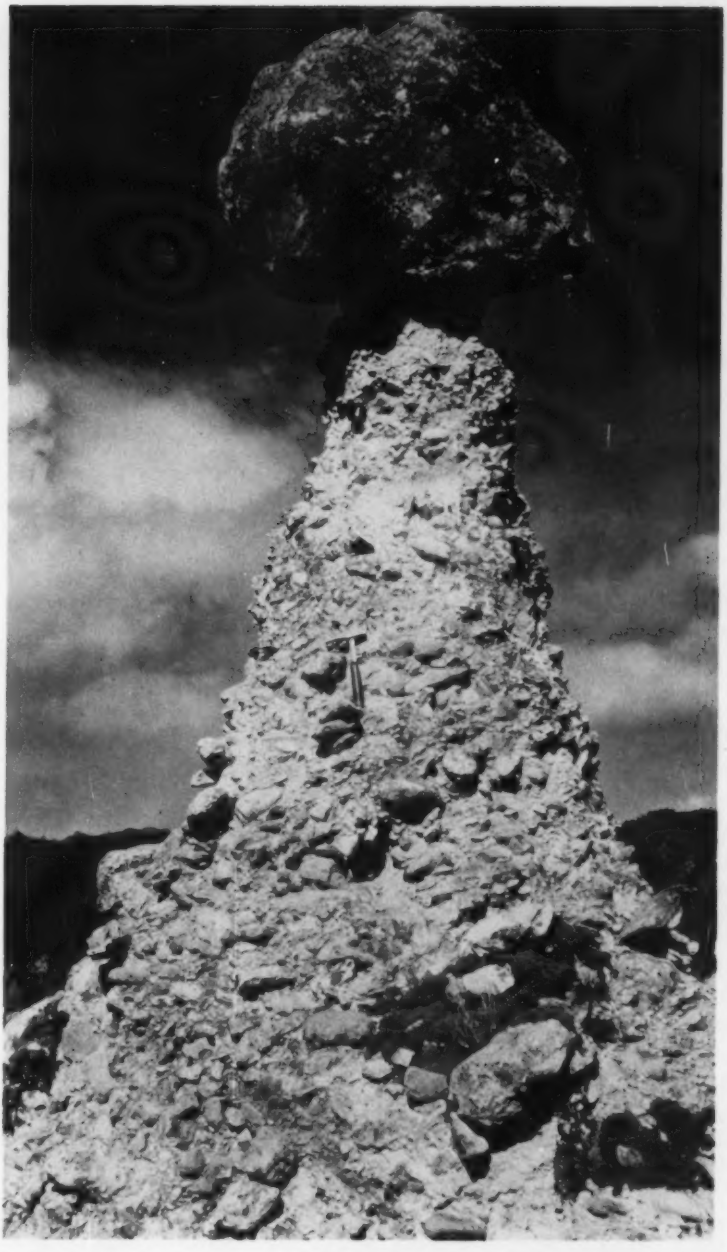
A

B

A, Large boulders of pre-Cambrian gneissic granite from Moncrief gravel on top of Bald Ridge.

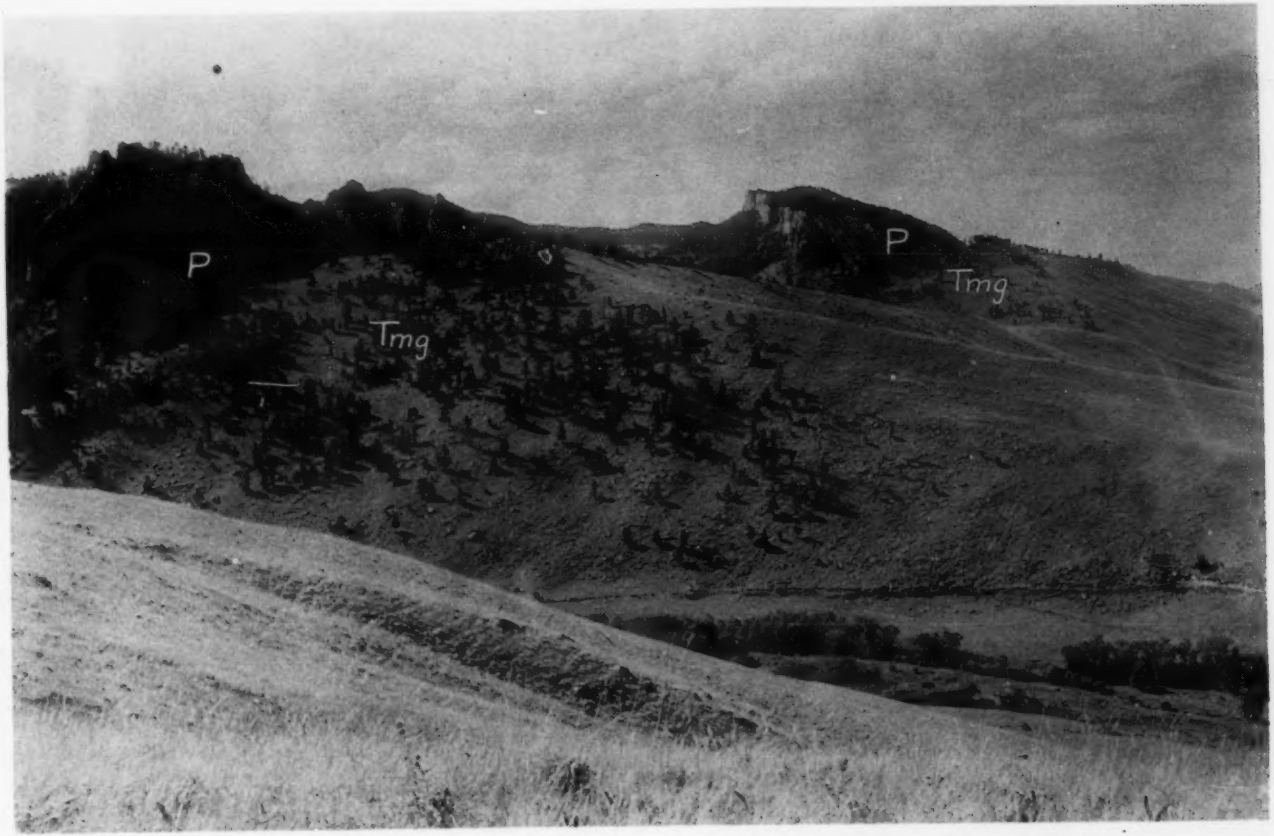
B, Looking southeast near mouth of Clear Creek with Bald Ridge on left skyline. Moncrief gravel in left half, separated from Paleozoic beds of mountain front by a west-dipping thrust fault.

PLATE III



Perched boulder formed by weathering in fanglomeratic phase of Kingsbury conglomerate, north fork of Rock Creek.

PLATE IV



View northward at mouth of French Creek showing Moncrief gravel (*Tmg*) abutting against Paleozoic strata (*P*)

(1941, pp. 167–168), although not explicit, seems to represent the same view. In 1909 J. A. Taff (p. 131) described the gravel composing Moncrief Ridge as gradational into the fine-grained early Tertiary coal-bearing beds of the Sheridan coal field; but his contribution seems to have escaped the notice of subsequent workers. Taff's views and, with some modification, those of Demorest approach most closely the interpretation favored in this study.

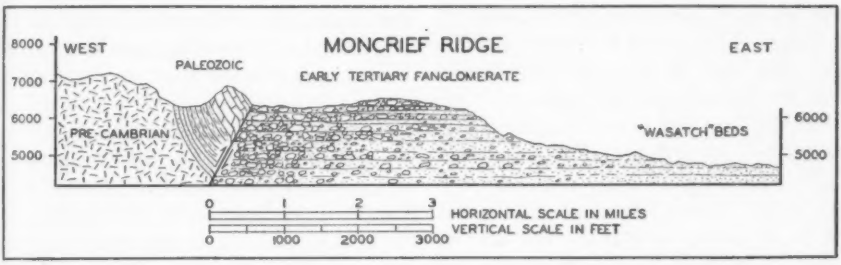


FIG. 4.—Sketch showing gradation of early Tertiary fanglomerate (Moncrief gravel) downward and eastward into "Wasatch" beds.

DESCRIPTION OF THE MONCRIEF GRAVEL

LITHOLOGY AND CONSTITUTION

This deposit features beds composed of subangular to rounded boulders of pre-Cambrian granitic rocks 1–5 feet in diameter; and larger stones are usually present. For example, on Moncrief Ridge are boulders up to 15 feet long; north of Little Piney Creek (pl. 1, *B*); and between Johnson and French creeks are boulders 25 feet long; south of Little Piney Creek one boulder was measured at 20 × 12 × 10 feet; and on the west end of Bald Ridge is one measuring 27 × 22 × 10 feet. Numerous boulders 10–15 feet in diameter are not exceptional at most exposures (pl. 2, *A*). These larger stones are subrounded, and most are composed of pre-Cambrian granite and gneissic granite. They lie as much as 1 mile east of the mountains and an additional mile from the nearest exposure of pre-Cambrian bedrock. Other lithologies represented, but in smaller amounts, are pre-Cambrian pegmatite, diabase, hornfels, gneiss, and schist, all of which are exposed in the core of the Big Horn Range. Stones derived from Paleozoic formations are extremely rare in the higher beds, but boulders of Paleozoic limestone compose up to 15 per cent of the lower strata, 900–1,400 feet below the top of Moncrief Ridge. The limestone boulders are subrounded to rounded and 6–18 inches in diameter at the most.

Matrix in the gravel is predominantly arkosic and usually sparse, particularly in the upper part where the boulder beds are coarsest, thickest, and most numerous. However, even here fines are not entirely lacking, for beds of arkose and fine micaceous siltstone outcrop in the midst of great boulder layers within 155 feet of the top of Moncrief Ridge (pl. 1, *A*). The deposits become finer downward and eastward, with an increase in number and thickness of arkose and micaceous arkosic sandstone layers. Boulder beds become fewer and thinner and their constituents smaller in the same directions

(fig. 4); and near the base cobble and pebble beds are common.

The finer materials contain crudely spherical arkosic concretions, 1–2 feet in diameter, formed by local cementation. These are identical with concretions in the so-called "Wasatch" of this area. The micaceous sandstone contains smaller limonitic concretions. Cementation of the gravel is poor except locally, where some arkose layers are well cemented and weather out in large fragments or outcrop as ledges. The gravel is white to light-gray and tan. Beds containing stones rich in ferromagnesian minerals are brownish, and micaceous sandy layers are colored grayish-green by chloritized biotite. Bedding is nearly indistinguishable in the coarser phases of the deposit (pl. 1, *A*), and at best it is crude and irregular with many scour channels and some cross bedding in the fine layers. Bits of lignitized woody material are included in fine beds low in the section,[2] and one decomposed lignitized log 1½ feet in diameter was found in the river bank behind the powerhouse on Clear Creek (NE. ¼ of SE. ¼ sec. 1, T. 50 N., R. 83 W.). Disintegration is so far advanced in many exposures that 50–90 per cent of the smaller stones can be chopped away with a geological pick, and some boulders up to 5 feet in diameter are friable to the core. Disintegration is most extensive in the upper 75 feet of the deposit.

Many features of the Moncrief gravel almost exactly duplicate S. H. Knight's (1937, p. 84), description of giant conglomerates at Green and Crook's mountains, Wyoming, and closely resemble conglomerates in the Wind River formation at the south end of the Big Horn Mountains (Tourtelot, 1946).

[2] Demorest (1938, p. 21) also reports coalified wood in these deposits.

THICKNESS

Thickness of the Moncrief gravel has been variously given as: probably exceeding 100 feet (Salisbury and Blackwelder, 1903, p. 22); at least 300 feet and possibly several hundred feet (Darton, 1906*a*, p. 69); 700–800 feet (Alden, 1932, p. 42); and more than 1,000 feet (Taff, 1909, p. 131). In this study, 485 feet of coarse bouldery gravel were measured by hand-leveling in continuous exposures on the north face of Moncrief Ridge (pl. 1, *A*), and similar boulder beds outcrop at least 900 feet below the top of the ridge on its north side. Along the south side of Moncrief Ridge, gravel beds with pre-Cambrian boulders 2½ feet in diameter outcrop along the bank of North Piney Creek, 1,400 feet vertically below the highest exposure of similar material. New cuts on U.S. Highway 16 along Clear Creek west of Buffalo show that the coarse boulder beds composing North Ridge are at least 1,200 feet thick.[3] It is safe to say that, close to the mountains, gravel composed predominantly of pre-Cambrian boulders is at least 1,200 feet thick on Clear Creek and 1,400 feet thick at Moncrief Ridge.

DISTRIBUTION

The largest accumulations of Moncrief gravel are in Bald and North ridges, 5 miles west of Buffalo, and in Moncrief Ridge, 2½ miles northwest of Story. However, as shown in figure 3, other sizable deposits of the gravel lie along the base of the range near French, Johnson, Shell, and the several branches of Piney creeks.

Boulders, up to 5 feet in diameter, of pre-Cambrian gneissic granite on Kingsbury Ridge, 4 miles south-southeast of Bald Ridge and 3 miles from the moun-

[3] Demorest (1938, p. 21) recognized that the gravel composing North Ridge was about this thick but gave no figures.

tains, have been mapped by Darton (1906*a*, pl. 47; 1906*b*, areal geology map) and Demorest (1938, fig. 5, p. 18*a*) as part of the Moncrief gravel and are so shown in figure 3. However, this occurrence is regarded with some skepticism because actual exposures of the gravel could not be found and large boulders of Bighorn dolomite (Ordovician) are also present. This material might be coarse debris washed out from Bald Ridge and the mountains during late Cenozoic degradation of the piedmont area.

In all other places Moncrief gravel lies directly against the upturned Paleozoic beds of the range front, which rise several hundred feet higher (pl. 2, *B*, and pl. 4). The gravel attains its greatest elevation, 6,900 feet, in Bald Ridge; and the lowest exposure is along North Piney Creek at 5,100 feet. It extends at least 2½ miles eastward from the mountain front, discounting the questionable occurrence on Kingsbury Ridge.

The subsummit upland of the Big Horn Mountains (fig. 2) is mantled with "Tertiary" gravels, which Demorest (1938, p. 21; ftn. 1) thought were to be correlated with the uppermost beds of the Moncrief. It is not unlikely that the Moncrief gravel formerly overlapped the lower flanks of the mountains, but the gravels on the subsummit upland are thought to be much younger, on the basis of structural and age relations to be detailed shortly. No deposits of the Moncrief were found on the subsummit upland.

Especially worthy of note are the map relations (fig. 3), which show that the Moncrief gravel occurs only where early Tertiary formations lie close to the base of the range. Where such formations are lacking, there is no gravel, and this holds even locally, as in Mowry Basin (fig. 3).

STRATIGRAPHIC AND STRUCTURAL RELATIONS OF THE MONCRIEF GRAVEL WITH THE "WASATCH"

Essentially horizontal beds of arkose and gray-green micaceous arkosic sandstone along the east front of the central Big Horn Mountains are shown on the latest maps (U.S. Geol. Survey, 1925, 1932) as "Wasatch,"[4] presumably on the basis of revisions (Taff, 1909, pp. 127–131; Wegemann, 1917, p. 60; Thom and Dobbin, 1924, pp. 494–498) of Darton's 1906*a*, p. 66; 1906*b*, p. 8) original treatment, in which they were dated as Tertiary or Cretaceous. Layers of identical material are interbedded with the Moncrief gravel even in its highest and coarsest part, and similar fine materials become more abundant downward and eastward in the section. This relation is particularly clear on Moncrief Ridge, where there can be little doubt that the gravel is but a coarse phase of the so-called "Wasatch"[5] (fig. 4). The boulder beds can be seen to lens out eastward along North Piney Creek, so that exposures of the "Wasatch" at the same level several miles east of the mountains consist principally of coarse arkose and micaceous arkosic sandstone. Fine beds in the Moncrief gravel contain coalified plant debris, a further similarity with the "Wasatch," which is coal-bearing in the Sheridan and Buffalo coal fields.

Previous workers, except Taff (1909, p. 131) and Demorest (1941, p. 168; ftn. 1),

[4] R. L. Nace (1936, p. 121) suggests that the name "Wasatch" be avoided as unnecessary and meaningless. Proper identification and naming of Paleocene and Eocene units east of the Big Horn Mountains remains for future detailed paleontological and stratigraphic studies, and the word "Wasatch" as used here is simply a quotation from earlier maps and publications.

[5] Taff (1909, p. 131) recognized and properly interpreted this relation in 1909.

have inferred or postulated an unconformity between the gravel and the "Wasatch," but extended search failed to reveal any discernible break between these units.

WITH THE KINGSBURY CONGLOMERATE

The Kingsbury is a limestone conglomerate, locally fanglomeratic, defined and named by Darton (1906*a*, p. 60) from Kingsbury Ridge (fig. 3). As he mapped the formation, it is a lens-shaped deposit, 2–5 miles wide, extending 40 miles along the east base of the central Big Horns. Wegemann (1917, p. 59) speaks of it as a "fan deposit," and Darton (1906*b*, p. 8) states that it was produced by uplift of the range. Demorest (1938, p. 20) also emphasizes that the Kingsbury is a fanglomerate, formed by uparching of the central segment of the Big Horn Mountains.

One of the best exposures of the Kingsbury is in prominent bluffs northeast of the North Fork of Rock Creek (sec. 19, T. 52 N., R. 83 W., Fort McKinney quadrangle), where it is chiefly a heterogeneous, poorly bedded, massive fanglomerate of subangular to subrounded fragments of Paleozoic limestone up to 5 feet in diameter (pl. 3). No boulders of pre-Cambrian origin were found here, although a few have been seen in other exposures. Cementation, at least locally, is good. Not less than 400 feet of beds are exposed north of Rock Creek; and Darton (1906*b*, p. 8) gives the maximum thickness of the formation as 2,500 feet (fig. 5).

In other localities farther from the mountains, the Kingsbury is predominantly a conglomerate containing well-rounded to subrounded pebbles and cobbles of Paleozoic limestones and occasional stones of pre-Cambrian origin. The Kingsbury is unconformable on tilted Mesozoic beds (Knowlton, 1909, p. 209; Gale and Wegemann, 1910, p. 144; Wegemann, 1917, p. 60; Demorest, 1938, p. 19), although Darton (1906*b*, p. 8) thought this contact was conformable and dated the formation as later Cretaceous. Wegemann (1917, p. 60) placed the Kingsbury in the "Wasatch," but Knowlton (1909, p. 210) later summarized fossil evidence indicating that it is probably Fort Union, and it is so shown on the latest maps (U.S. Geol. Survey, 1925, 1932). Nace (1936, p. 100) includes the Kingsbury in the "Wasatch Group" for purposes of discussion and summarizes the conflicting evidence as to its age, which he tentatively puts as Paleocene(?); but Jepsen (1940, p. 242) cites fossil evidence indicating that the Kingsbury is Early Eocene. Roland W. Brown, of the United States Geological Survey, has collected vertebrate remains from the Kingsbury conglomerate which definitely fix its age as Eocene.[6]

Bedding in the Moncrief gravel, where discernible, is essentially horizontal, with eastward dips not exceeding 2°–3°. All dips measured in the Kingsbury conglomerate are 17°–23° northeastward. The actual contact between the Moncrief and the Kingsbury has not been seen; but, because of this discordance in dips, it is thought to be an angular unconformity, and the marked difference in composition of the two formations also suggests a break between them (fig. 5). This was Darton's (1906*b*, p. 8) original interpretation, as well as Alden's (1932, p. 42); but Demorest (ftn. 1; 1941, pp. 167–168) described this contact on the north side of Moncrief Ridge as gradational, although in earlier work (1938, p. 21) he treated the Moncrief-Kingsbury contact in the area of Clear Creek

[6] Personal letter (1947).

as an angular unconformity. It now appears that the gradation described by Demorest is the Moncrief-"Wasatch" relation treated in the preceding pages.

In places, outcrops of the Kingsbury conglomerate have greater elevations than the lowest near-by exposures of Moncrief gravel. This relation is well shown on Little Piney Creek and is thought to indicate that the Moncrief was deposited on a land surface of considerable local relief (fig. 5). Ridges and hills of this early landscape, held up by firmly cemented Kingsbury conglomerate, have been exhumed in the modern cycle of erosion.

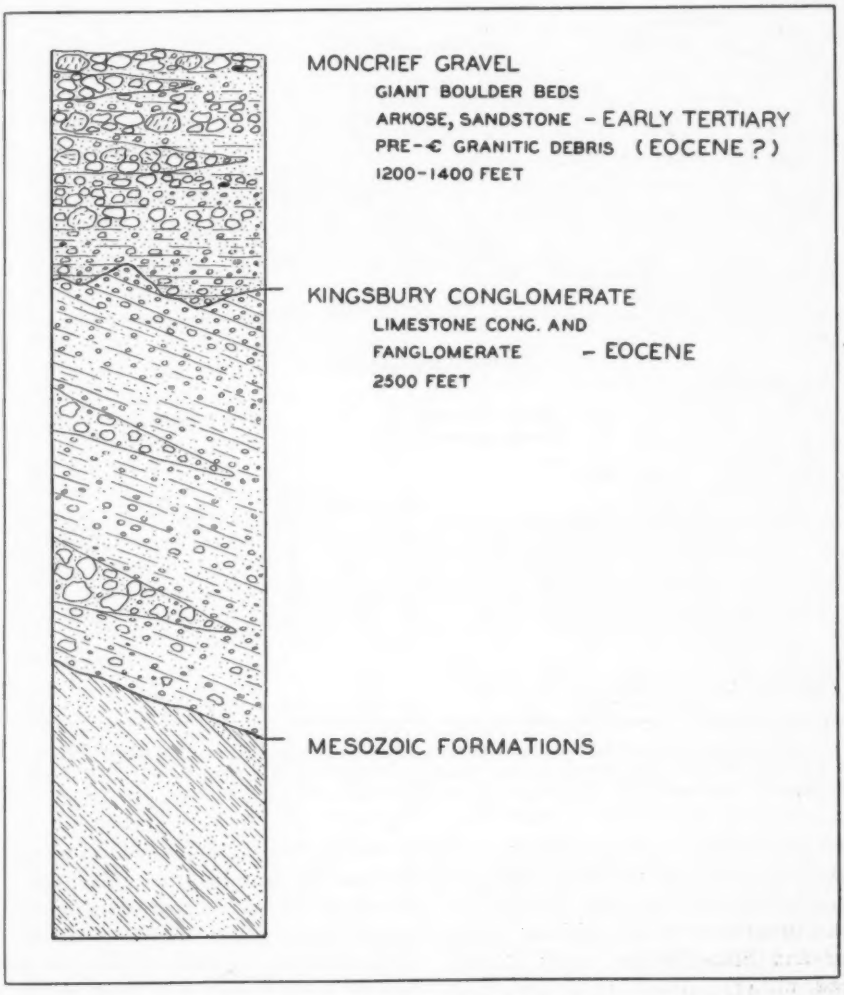


FIG. 5.—Columnar section of Moncrief gravel and Kingsbury conglomerate

WITH PRE-TERTIARY ROCKS

The Moncrief gravel rests on tilted Mesozoic strata with a marked angular unconformity, as recognized by Darton (1906*a*, pl. 47; 1906*b*, areal geology map) and as clearly demonstrated by the field pretation meets insurmountable obstacles. One has but to see the great coarse-boulder layers, composed wholly of pre-Cambrian rock fragments, abutting directly against the upturned Paleozoic limestone beds (pl. 2, *B*, and pl. 4)

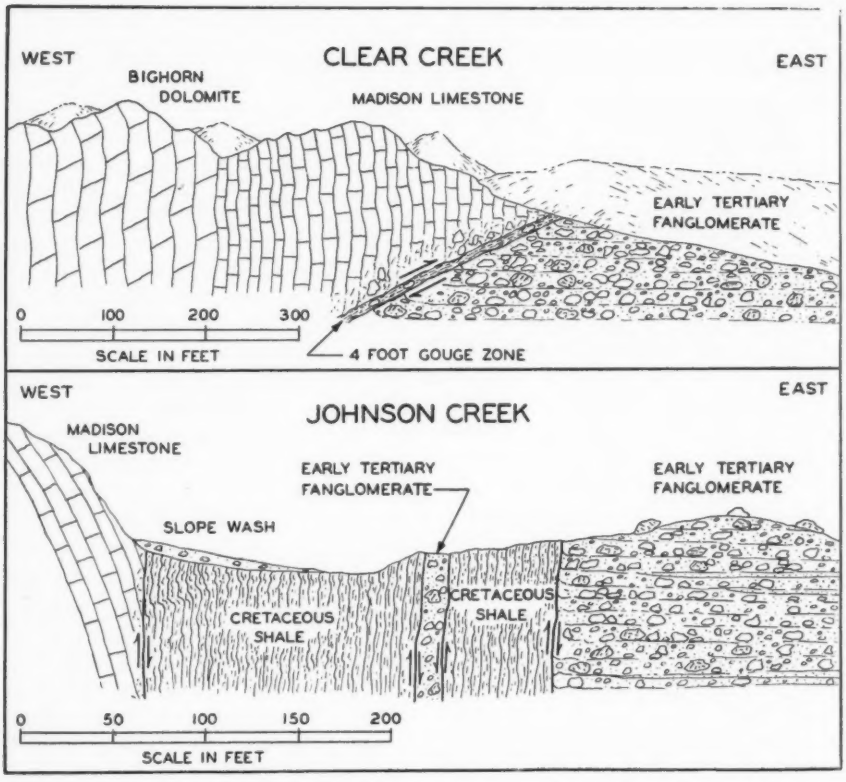


FIG. 6.—Field sketches of fault relations between early Tertiary fanglomerate (Moncrief gravel) and pre-Tertiary rocks of the Big Horn Mountains. Exposure at Clear Creek is in a road cut on U.S. Highway 16. The Johnson Creek exposure is in a gully on the first spur to the north.

and map relations (fig. 3). The contact between gravel and upturned Paleozoic beds forming the mountain front has also been interpreted as depositional (Salisbury and Blackwelder, 1903; Darton, 1906*a*, pl. 47; Alden, 1932; Demorest, 1941, pls. 1 and 4); but such an interto realize that the Paleozoic rocks could not possibly have had their present position when the gravel was being deposited, unless the gravel came from the east. This possibility is most remote, for the nearest known source of pre-Cambrian debris to the east is the Black Hills, 145

miles away. The relations are those expected of a fault; and actual exposures of a fault contact between Moncrief gravel and Paleozoic rocks have been observed at Clear Creek, north of Johnson Creek, and on North Piney Creek. Details of the fault exposures on Clear and Johnson creeks are sketched in figure 6. In the summer of 1946 a new road cut on U.S. Highway 16 along Clear Creek gave a superb exposure of the fault, showing Paleozoic limestone thrust eastward over Moncrief gravel along a plane dipping 25° west. On Johnson Creek the fault is a complex zone, consisting of several nearly vertical planes of displacement involving Paleozoic limestone, distorted Cretaceous shale, and the Moncrief gravel. Before faulting, the gravel probably overlay the Cretaceous shale with an angular unconformity; and in the subsequent displacement a thin sliver or wedge of the gravel was dropped down into the shale, which, in turn, forms a larger slice between the Paleozoics and the gravel. The fault planes appear nearly vertical in the exposure, but they may dip westward at depth. On North Piney Creek the fault is a thrust dipping westward.

On the basis of these exposures and from the fact that Paleozoic strata are steeply tilted or even overturned wherever the Moncrief gravel abuts against them, the contact is interpreted as a fault in all places. Nowhere was a depositional contact observed between these two units. Stratigraphic throw on the fault must have been at least 2,000 feet to give the present relations. Thrust faults have long been recognized (Darton, 1906*b*; Demorest, 1941; Bucher, Thom, and Chamberlin, 1934, p. 169) along the east base of the central Big Horns, but it has not been shown previously that the Moncrief gravel is involved in some of the thrusting.[7] Faulting occurred at more than one period, for Demorest (1941, pp. 165–166) and Darton (1906*a*, p. 93) recognize at least two episodes of faulting. Darton (1906*b*, pp. 60–61, 93) also speaks of the Kingsbury conglomerate–Paleozoic contact as being a fault in some places and an overlap elsewhere. Some skepticism is entertained with respect to a few of Darton's fault relations, as he appears to have mapped great landslide masses as fault blocks.

AGE OF THE GRAVEL

The Moncrief gravel has previously been dated as early Pleistocene (Salisbury and Blackwelder, 1903, pp. 222–223; Alden, 1932, p. 42), late Tertiary or earliest Quaternary (Darton, 1906*a*; 1906*b*), and has been compared with the Bishop conglomerate (Atwood and Atwood, Jr., 1938, p. 966; Rich, 1910, pp. 601–632) of the Uinta Mountain region, which is thought to be Miocene or Oligocene by W. H. Bradley (1936, p. 163, 185) or about Pliocene by W. W. Atwood (1940, p. 313). Demorest (1941, pp. 167–168) includes the Moncrief beds in his Tertiary(?) gravels, which may range from Eocene well up into the Tertiary. Taff (1909, p. 131) thought the gravel at Moncrief Ridge was equivalent to nearly all his upper member in the Sheridan coal field, which includes the Tongue River member of the Fort Union (Paleocene) and the Intermediate and Ulm coal groups, which may be Eocene (Thom and Dobbin, 1924, pp. 494–498;

[7] The 1947 Bighorn Basin field conference guidebook prepared by the University of Wyoming, the Wyoming Geological Association, and the Yellowstone-Bighorn Research Association, which appeared since this paper was prepared for publication, makes note (p. 95) of the fault between Moncrief gravel and Paleozoic rocks exposed on U.S. Highway 16 along Clear Creek.

Baker, 1929, p. 24; Nace, 1936, pp. 91, 95, 104).

An early Tertiary age is favored in this paper because the gravel appears to be part of the "Wasatch" on two counts. First, it contains typical "Wasatch" material, including coalified plant remains, and passes by gradation downward and eastward into "Wasatch" beds. Second, the gravel occurs only where the "Wasatch" lies along the base of the range. Structural relations also suggest an early Tertiary age, for the gravel is older than some of the eastward thrusting of the Big Horn Mountains, which is presumably Laramide. If the gravel were early Pleistocene, we should have evidence of at least 2,000 feet of Pleistocene thrust-fault displacement—a new and unlikely chapter in the evolution of the Middle Rocky Mountains.

Exact dating of the Moncrief gravel awaits detailed paleontologic and stratigraphic study of early Tertiary beds in the Sheridan coal field. As it now stands, Taff's (1909, p. 131) original description, as modified by more recent work (Thom and Dobbin, 1924; Baker, 1929), makes the lower part of the Moncrief gravel Paleocene and the upper part Eocene; but the evidence of angular unconformity between Moncrief gravel and Kingsbury conglomerate, already described, renders a Paleocene age for any part of the Moncrief unlikely. The resemblance to conglomerates described by Tourtelot (1946) in the late lower Eocene Wind River formation at the south end of the Big Horn Range strengthens the probability that the Moncrief gravel is Eocene.

ORIGIN

Proponents (Salisbury and Blackwelder, 1903; Alden, 1932) of a Pleistocene glacial origin for the Moncrief gravel have based their arguments chiefly on relations at Bald, North, and Moncrief ridges. However, the equally coarse gravel near Johnson, Shell, and Little Piney creeks is part of the same deposit, and it is noteworthy that none of these latter streams shows the slightest evidence of glaciation in their lower parts. Furthermore, neither Johnson Creek nor Little Piney Creek extends far enough into the range to reach areas of known glaciation (Darton, 1906*a*, pl. 26; 1906*b*, areal geology map). A Pleistocene age is further ruled out by stratigraphic and structural relations already detailed. These relations likewise render untenable Darton's (1906*a*, pp. 69–70; 1906*b*, p. 9) interpretation that the boulder deposits are late Tertiary or earliest Quaternary bench gravels. In this study the present outcrops of Moncrief gravel are interpreted as remnants of early Tertiary alluvial fans built up along the east base of the Big Horn Mountains as they were being uplifted during the Laramide Revolution.[8]

The composition of the Moncrief gravel clearly indicates derivation from the pre-Cambrian core of the Big Horn Mountains; and the crude bedding, angularity, and size of boulders, sparseness of matrix, and occasional fine beds are all consistent with fan deposition (Trowbridge, 1911, pp. 706–747). The huge

[8] This concept was outlined by Professor W. T. Thom, Jr., in a conversation in 1940; and Demorest had essentially the same idea. It was viewed with skepticism during the early stages of this work, until the accumulating field evidence rendered other interpretations untenable. Love (1940) and Tourtelot (1946) report boulder beds in lower Eocene deposits along the south flank of the Big Horn Mountains; and coarse fanlike deposits in early Tertiary beds bordering other Middle Rocky Mountain ranges have been described by the following: Eldridge (1894, pp. 25–26); Westgate and Branson (1913, p. 144); Blackwelder (1915, pp. 106–108); Bauer (1934, pp. 678–679); Love (1939, pp. 60, 106–107); Branson and Branson (1941, pp. 142–143); Van Houten (1944, pp. 179–181).

boulders furnish no objection, for equally large boulders are found on alluvial fans at the east base of the Sierra Nevada (Trowbridge, 1911, pp. 716–717, 740–743). Conditions causing fan deposition were presumably brought about by uplift of the mountains, and this uplift was probably by stages.[9] Initially Mesozoic and Paleozoic strata were exposed in the uparched area; and deposits composed primarily of Paleozoic fragments, such as the Kingsbury conglomerate, were formed along the flanks of the range. Darton (1906*b*, p. 8), Wegemann (1917, pp. 59–60), and Demorest (1938, p. 20; 1941, p. 168) all recognize the Kingsbury as a product of mountain uplift. A subsequent uplift which extended into the piedmont area deformed the Kingsbury mildly and initiated erosion. This was followed by deposition of coarse fan materials, the Moncrief gravel, derived from the mountains which were being raised abruptly above the piedmont, presumably by faulting. At first, the fans consisted of mixed Paleozoic and pre-Cambrian rock fragments; but, as stripping of the Paleozoic rocks became more complete, a larger percentage of the debris was derived from the pre-Cambrian core, until, finally, practically all the detritus was of pre-Cambrian origin.

Increase in coarseness of the gravel upward in the section may have been due to the more massive nature of the pre-Cambrian rocks at depth, to an increasing rate of uplift, or possibly to early Tertiary glaciation in the high parts of the range, like that recognized elsewhere in the Rocky Mountains (W. W. Atwood, 1915, pp. 13–26; W. W. and W. R. Atwood, 1926, pp. 612–622; W. W. and W. W. Atwood, Jr., 1938, p. 961; Drysdale, 1915, pp. 65–66; Scott, 1938, pp. 628–636). C. J. Hares (1926*a*, pp. 174–175; 1926*b*, p. 175; 1926*c*, pp. 175–176) has suggested that somewhat similar deposits along the flanks of various Rocky Mountain ranges are the product of a widespread mid-Tertiary glaciation, but this has apparently not been confirmed by subsequent investigators. The Moncrief gravel is certainly not a till; but it may consist of coarse outwash debris, although there is no direct evidence indicating a glacial source. Alden (1932, p. 43) found one boulder with something like chatter marks; and one small stone with striae-like scratches was collected during the present investigation. An early Tertiary glacial source for the gravel is not established by such meager evidence, and it is not essential to the major thesis proposed. In fact, the map relations (fig. 3) showing that the Moncrief gravel is best developed along those parts of the range uplifted by thrust faults suggest that the gravel was derived from rising thrust blocks as proposed by Knight (1937) and Tourtelot (1946) for similar gravels elsewhere in Wyoming. The fault contact between Moncrief gravel and pre-Tertiary rocks indicates that movement on the faults also occurred after deposition of the gravel.

[9] Love (1939, pp. 116–117) recognizes eight pulsations of the Laramide Revolution in western Wyoming.

CONCLUSIONS

A series of coarse bouldery fans composed primarily of pre-Cambrian debris was built up along the east base of the central Big Horn Mountains in early Tertiary time, probably the Eocene. This was the section of greatest uplift, and the fan debris was presumably coarser and thicker here than elsewhere. Subsequent faulting thrust the Paleozoic beds of the mountain front eastward against the gravel, and erosion during the remainder of the Cenozoic has gradually etched out

the finer materials, leaving the thickest and coarsest parts of the fan deposits as prominent ridges in the present landscape. At least three episodes of Laramide deformation are indicated: (1) An uplift which produced the Kingsbury conglomerate. Demorest (1941, p. 168) cites evidence indicating that faulting probably occurred during this uplift. (2) A second uplift, also probably attended by faulting, which deformed the Kingsbury and produced the coarse Moncrief gravel. This episode may have been accompanied or closely followed by alpine glaciation. (3) A third, post-Moncrief, period of thrust faulting toward the east in the central segment of the range.

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PHYSIOGRAPHIC DIVISIONS OF ILLINOIS

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ABSTRACT

The classification proposes minor modifications in Fenneman's divisions and recognizes subdivisions of the Till Plains and Great Lakes sections, based largely on glacial features. The boundaries and characteristic features of the subdivisions are described, and their origin and relations to glacial features and bedrock topography are discussed.

INTRODUCTION

During the last sixty years increasing attention has been given to physiographic classification; and the broad outlines first conceived by W. M. Davis, J. W. Powell, and others have crystallized into a widely accepted classification for the United States. The establishment of this classification was due largely to comprehensive studies by the late N. M. Fenneman (1914, pp. 84–134; 1928, pp. 261–353; 1931; 1938). In these studies it was recognized that, with the progress of topographic and geologic mapping and the advances in geomorphic knowledge, numerous refinements and revisions would be made. It is the purpose of the present report to make such adjustments of Fenneman's regional boundaries in Illinois as are warranted by present information and to delineate smaller subdivisions which distinguish physiographic differences that can be shown on large-scale maps and which can be used in regional studies within the state (table 1 and fig. 1). These subdivisions, in turn, may be broken down later into still smaller units which may be used as a basis for description in quadrangle reports and other local studies. Local physiographic areas of this type have been described by F. M. Fryxell (1927, pp. 1–53) and by H. B. Willman (1942, pp. 31–37). In the present paper only the distinctive characteristics of the larger subdivisions are summarized.

The report is the outgrowth of geomorphic and glacial investigations carried on by Leighton since 1919, by Ekblaw since 1923, and of recent studies of the bedrock topography and subsurface Pleistocene deposits by Horberg. Copies of a preliminary draft of the map of physiographic divisions of Illinois were prepared in 1944 and furnished to the Committee on Drainage Basins and Flood Control of the Illinois Post-war Planning Commission and to the Illinois Legislative Flood Investigating Commission.

GENERAL DESCRIPTION AND REGIONAL RELATIONS

Illinois is essentially a prairie plain, and, compared with many other states, it presents few striking physiographic contrasts. The relief over most of the state is moderate to slight and is not sufficient to exert a marked effect on climate. Situated in the south-central part of the great Central Lowland (fig. 2) and near the confluence of major lines of drainage, it is the lowest of the north-central states. The mean elevation is about 600 feet above sea-level, compared with 700 feet for Indiana, 1,050 feet for Wisconsin, 1,100 feet for Iowa, and 800 feet for Missouri (Gannett, 1892, p. 289). The total relief of the state is 973 feet, the highest point, 1,241 feet above sea-level, being Charles Mound in the northwest corner of the state, and the

lowest point, 268 feet above sea-level, the junction of the Ohio and Mississippi rivers. The greatest local relief is near the major valleys, especially within the driftless uplands of northwestern and southern Illinois, and reaches a maximum of 775 feet between Williams Hill, 1,065 feet above sea-level, and the Ohio River Valley, 290 feet above sea-level, in Pope County of southern Illinois. The local relief in most counties in the state, however, is less than 200 feet.

Although large-scale relief features are absent, the physiographic divisions of the state are readily apparent and assume great local significance. More than nine-tenths of the state lies within the Central Lowland, all of which is glaciated except the Wisconsin Driftless section in northwestern Illinois. The other provinces—Ozark Plateaus, Interior Low Plateaus, and Coastal Plain—lie almost entirely outside the glacial boundary in southern and southwestern Illinois.

TABLE 1

PHYSIOGRAPHIC CLASSIFICATIONS OF ILLINOIS

Classification by Fenneman	Classification by Leighton, Ekblaw and Horberg
Central Lowland province	Central Lowland province
Great Lake section	Great Lake section
	Chicago Lake Plain
	Wheaton Morainal Country
Till Plains section	Till Plains section
	Kankakee Plain
	Bloomington Ridged Plain
	Rock River Hill Country
	Green River Lowland
	Galesburg Plain
	Springfield Plain
	Mount Vernon Hill Country
	Dissected Till Plains section
Wisconsin Driftless section	Wisconsin Driftless section
Ozark Plateaus province	Ozark Plateaus province
	Lincoln Hills section
Salem Plateau section	Salem Plateau section
Interior Low Plateaus province	Interior Low Plateaus province
Shawnee section	Shawnee Hills section
Coastal Plain province	Coastal Plain province

FACTORS DETERMINING PHYSIOGRAPHIC CONTRASTS

The physiographic contrasts between various parts of Illinois are due to the following factors and conditions: topography of the bedrock surface; extent of the several glaciations; differences in glacial morphology; differences in age of the uppermost drift; height of the glacial plain above main lines of drainage; glaciofluvial aggradation of basin areas; and glaciolacustrine action.

TOPOGRAPHY OF THE BEDROCK SURFACE

Recently acquired knowledge of the topography of the bedrock surface of Illinois, so widely obscured by the glacial

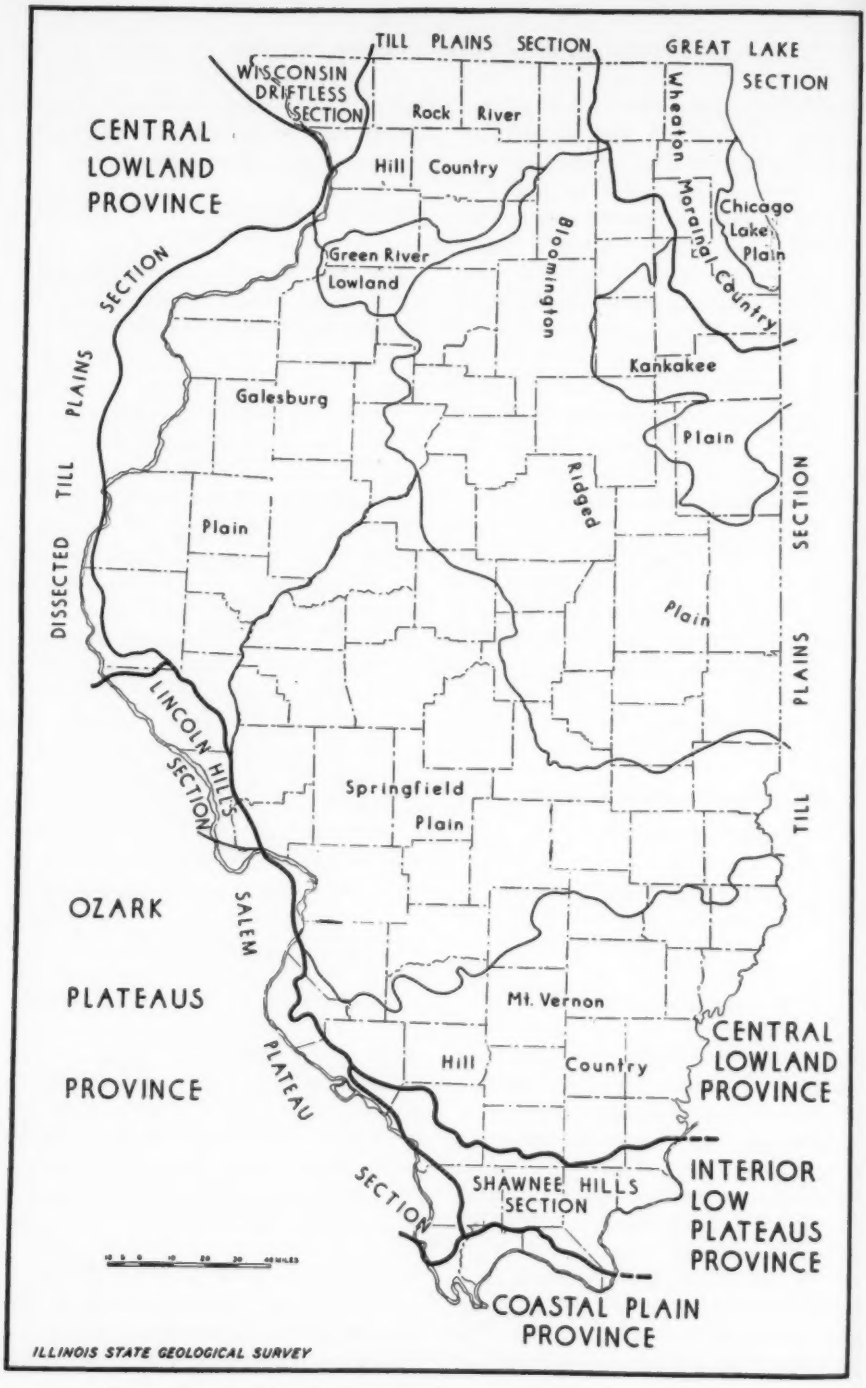

FIG. 1.—Physiographic divisions of Illinois, by Leighton, Ekblaw, and Horberg

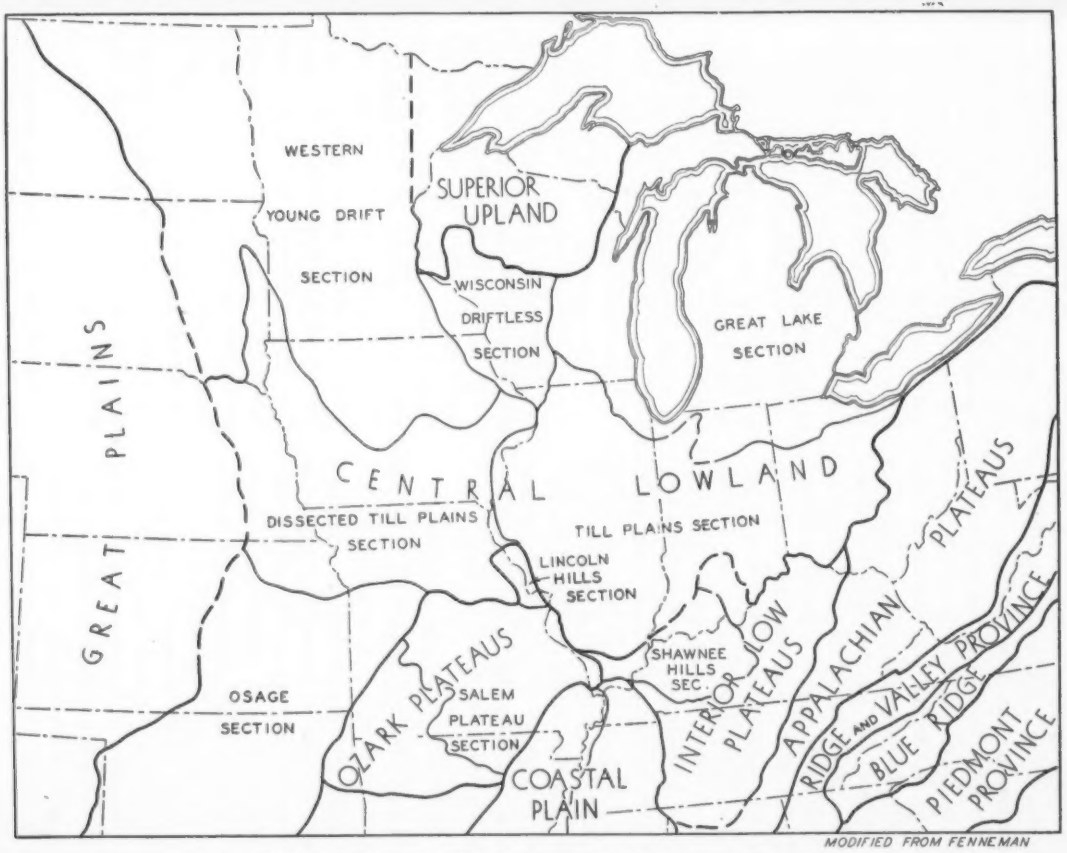

FIG. 2.—Major physiographic divisions in central United States (after Fenneman)

deposits, emphasizes the great importance of this factor in the shaping of the broad physiographic features of the state (Horberg, 1946, pp. 179–192). Prior to glaciation an extensive lowland, eroded on the weak Pennsylvanian rocks of the Illinois basin, covered most of central Illinois. Bordering it on the north, west, and south were uplands that had been developed, for the most part, on the more resistant older Paleozoic limestones and dolomites. This ancient pattern is reflected to an important degree in the present land surface. The extensive lowland of central Illinois provided conditions for the thickest accumulation of glacial deposits and the development of the prairie plains of this portion of the state. The higher uplands of northwestern Illinois and of southern Illinois (the Shawnee Hills) prevented the further movement of the potent Illinoian glacial lobe and caused striking physiographic juxtapositions.

EXTENT OF THE SEVERAL GLACIATIONS

The effect of the extent of several glaciations is most apparent in the physiographic contrasts between the glaciated and nonglaciated areas of the state—the Wisconsin Driftless section and the bordering Rock River Hill Country in the northwest and the Shawnee Hills section and the Mount Vernon Hill Country in the south. This emphasis upon glaciation, however, should be modified by that measure of difference which already existed in the preglacial landscape. Even so, there is no denying the fact that a topographic revolution resulted from glaciation.

The successive superposition of younger drift-sheets upon older in some parts of Illinois and the deposition of only one drift-sheet in other parts also produced physiographic contrasts in the present topography (fig. 3). The bedrock control that characterizes the Mount Vernon Hill Country and most of the Rock River Hill Country is due to the fact that these regions have but one glacial mantle (the Illinoian), whereas the remainder of the glaciated area of the state was buried beneath either two or three drift-sheets. Nebraskan, Kansan, and Illinoian drifts are present in western Illinois, the first two from the Keewatin field; Kansan, Illinoian, Wisconsin, and possibly Nebraskan drifts from the Labradorean field are present in northeastern Illinois; and Kansan and Illinoian drifts (Labradorean) are present in west-central and south-central Illinois.

DIFFERENCES IN GLACIAL MORPHOLOGY

Two contrasting types of topography —the Wheaton Morainal Country and the Bloomington Ridged Plain—impressively illustrate differences in glacial morphology. When the Wheaton Morainal Country was formed, the ice lobe was more confined to the deep Lake Michigan basin, and the moraines are closely huddled together. In molding the Bloomington Ridged Plain, the glacier was more extensive and more widely radiating, and during its receding and readvancing substages it formed moraines widely spaced, alternating with nearly featureless ground-moraine plains. The moraines are also generally smoother than the bold moraines of the Wheaton Morainal Country.

DIFFERENCES IN AGE OF THE UPPERMOST DRIFT

Physiographic contrasts have also been produced by differences in age of the uppermost drift, as in the case of the Bloomington Ridged Plain compared with the Springfield Plain or the Gales-

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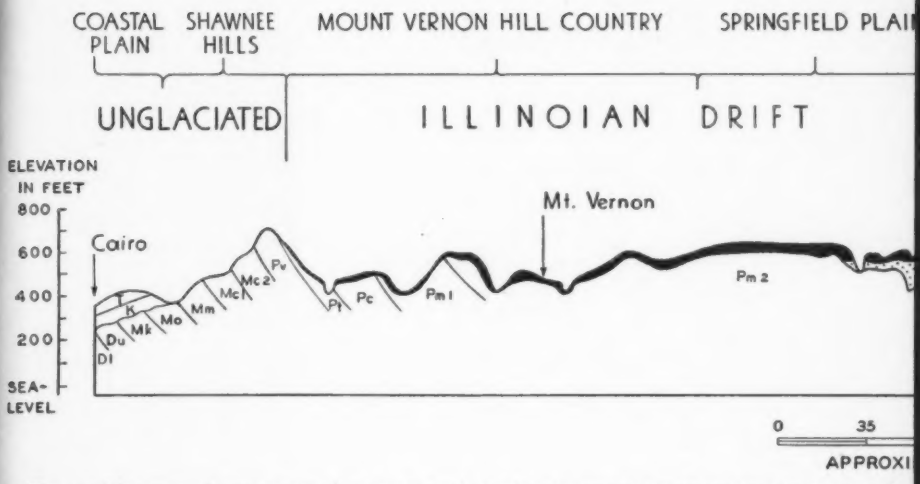
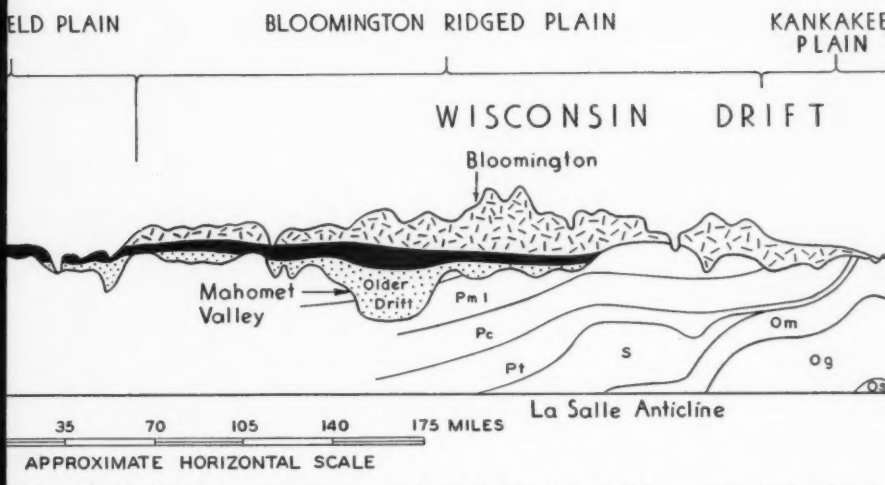


FIG. 3.—Diagrammatic cross section from Cairo north to Bloomington and thence northeast to Chicago, sho
bedrock formations. Note that the ratio of the vertical scale to the horizontal scale, 462: 1, exaggerates the dip o

KEY

T —Tertiary system
K —Cretaceous system
Pm2—McLeansboro group, upper part
Pm1—McLeansboro group, lower part
Pc —Carbondale group
Pt —Tradewater group
Pv —Caseyville g
Mc2—Chester grou
Mc1—Chester grou



Chicago, showing (1) physiographic divisions, (2) glacial deposits, (3) profiles of present and bedrock surfac
tes the dip of the bedrock formations.

KEY TO BEDROCK FORMATIONS

—Caseyville group	*Mm* Meramec group	*Du* —Devonian system, upper part	*Om* —Maquoket
—Chester group, upper part	*Mo* —Osage group	*Dl* —Devonian system, lower part	*Og* —Galena an
—Chester group, lower part	*Mk* —Kinderhook group	*S* —Silurian system	*Os* —St. Peter

NKAKEE WHEATON CHICAGO
PLAIN MORAINAL C'NTRY. LAKE PL.

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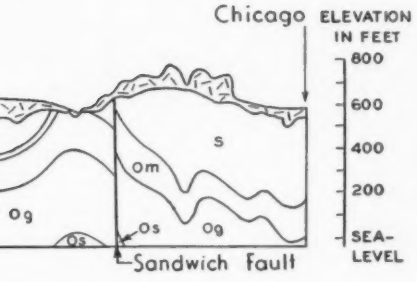

rock surfaces, and (4) succession and structure of

—Maquoketa formation
—Galena and Platteville formations
—St. Peter sandstone

burg Plain. The more youthful age of the Bloomington Ridged Plain is responsible for the preservation of its glacial landscape; the greater age of the other two is responsible for their erosional features.

HEIGHT ABOVE MAIN LINES OF DRAINAGE

The effect of height above main lines of drainage on physiographic contrasts is the basic factor in the differentiation of the Galesburg Plain from the Springfield Plain. The Galesburg Plain is sufficiently higher above the Illinois River so that its valleys are more sharply incised than were those of the Springfield Plain, and they have different stream regimens.

GLACIOFLUVIAL AGGRADATION OF A BASIN AREA

The Green River Lowland is an example of the glaciofluvial aggradation of a basin area.

GLACIOLACUSTRINE ACTION

The Chicago Lake Plain is an example of glaciolacustrine action.

PHYSIOGRAPHIC DIVISIONS

GREAT LAKE SECTION

The Great Lake section of the Central Lowland province is separated from the Till Plains section to the south because of the bold encircling moraines of Lake Michigan basin, the greater prominence of lakes, and the extent of lacustrine plains in this area. In Illinois a purely topographic boundary between the two sections was drawn by Fenneman along "the outer edge of certain late Wisconsin moraines," which, south of the Kankakee River, form "the western rim of the 'Kankakee swamp . . . .' " (1928, p. 315). In this report the boundary is drawn as follows: Along the west border of the Tazewell Marengo ridge (fig. 4) and the Hampshire ridge south from the Wisconsin-Illinois state line to a point in central Kane County where the Elburn moraine turns westward and southwestward, easterly along an arbitrary topographic boundary to the Cary Valparaiso moraine in western Du Page County, along the outer edge of the Valparaiso moraine southward to the Will County line, and along the outer edge of the closely related Rockdale-Manhattan moraine southward and southeastward to the Indiana-Illinois state line. This boundary coincides, in general, with Fenneman's. The only significant difference is in eastern Kankakee County, where the "Kankakee swamp" is excluded from the section, because the area is an alluviated glacial drainageway and lacks the characteristics of the true lacustrine plains found elsewhere in the section (Ekblaw and Athy, 1925, pp. 417–428). A similar revision of the boundary in the adjoining part of Indiana is implied.

The section is subdivided into the Wheaton Morainal Country and the Chicago Lake Plain, the boundary being marked by the highest (Glenwood) shoreline of glacial Lake Chicago.

Chicago Lake Plain.—The Chicago Lake Plain is the well-known featureless "prairie" of early writers of the area. It is characterized by a flat surface, underlain largely by till, which slopes gently lakeward and is interrupted by low beach ridges, morainic headlands and islands, and by two large glacial drainageways along the Des Plaines River and Sag Channel (Bretz, 1939, pp. 43–59). The beach ridges include well-developed spits and bars, which parallel the lake shore or funnel southwestward toward the outlet channel along the Des Plaines River. Blue Island, a prominent morainic island, is situated just east of the outlet channel and rises about 50 feet above

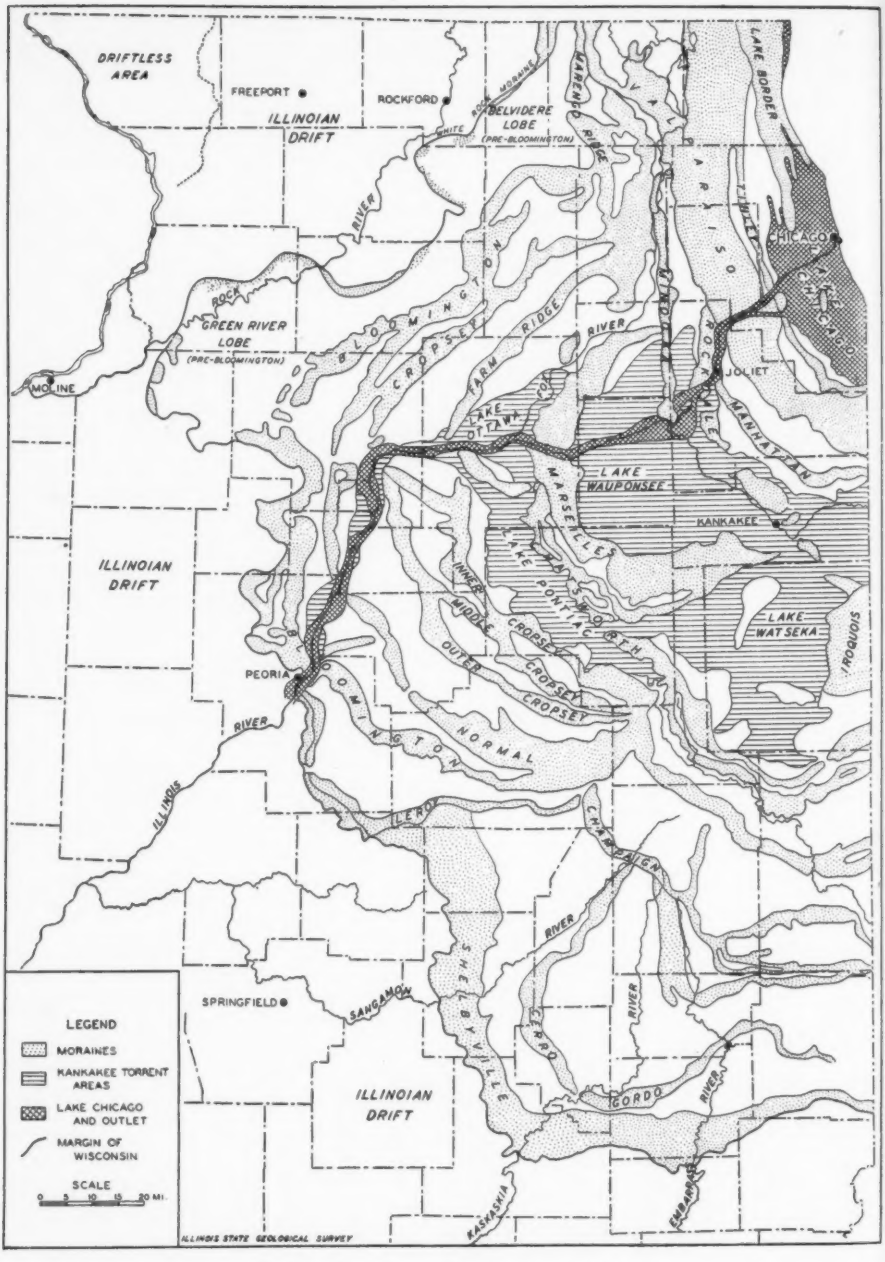

FIG. 4.—Glacial map of northeastern Illinois

the plain. Dunes, which are so conspicuous along the lake shore farther east, are scarcely recognizable and are found in only a few scattered localities.

Originally much of the lake plain was swampy and poorly drained. Its three rivers—the Chicago, the Calumet, and the Des Plaines—are without true valleys and have courses determined largely by beach ridges.

Wheaton Morainal Country.—The Wheaton Morainal Country is characterized by glacial morainic topography (mostly of the Cary substage), which is more complex in detail and has more lakes and swamps than do the open stretches of the adjoining Bloomington Ridged Plain. It includes a series of broad parallel morainic ridges, which encircle Lake Michigan. In detail the topography is complicated by a variety of elongated hills, mounds, basins, sags, and valleys. The area is dominated by the Valparaiso moraine, which has the highest elevation and, except where interrupted by valleys, is continuous from Wisconsin to Indiana. With the exception of the Tinley moraine, all other moraines are discontinuous geographic features—those in front of the Valparaiso moraine are overridden by it and those behind are either interrupted by the Chicago Lake Plain or merge with ground moraines. Kames, kame terraces, kettles, basins, and eskers, although not abundant, occur more commonly than elsewhere in the state. Fox Lake and associated lakes are conspicuous water bodies. Small basins of extinct lakes and ponds underlain by stratified silts and clays are found throughout the area.

The topography is determined essentially by thick Wisconsin drift of the Lake Michigan ice lobe, the deposits of which completely buried the underlying bedrock surface. Older Illinoian drift occurs below the Wisconsin deposits along the outer margin of the section in Kane and McHenry counties; but elsewhere this older drift appears to have been removed before the deposition of the Wisconsin drift (fig. 3).

Postglacial erosion has been slight and is restricted largely to youthful valleys along the Fox and Des Plaines rivers. Locally along the Des Plaines Valley near Lemont the Valparaiso drift-sheet is thin, and its major features are determined by a pre-Valparaiso valley system eroded into older underlying drift (Bretz, 1939, p. 52).

Geologic history.—The outer moraines in the Great Lake section, including the Marengo, Hampshire, and Cropsey, were formed during the Tazewell substage of the Wisconsin glaciation, whereas the Valparaiso, Tinley, and Lake Border moraines were deposited during the succeeding Cary substage. The Fox River Valley and the valleys of the east and west forks of Du Page River were trenched, apparently by torrential waters, before the advance of the Cary ice. During glacial recession, large blocks of stagnant ice became buried and, upon melting, left the large basins now occupied by Fox Lake and associated lakes.

As the Cary glacier withdrew into the basin of Lake Michigan, Lake Chicago was impounded behind the Tinley and Lake Border moraines, and its discharge waters eroded the outlet channel along the Des Plaines Valley. With retreat of the ice beyond the Straits of Mackinac, this lower eastward outlet into the Atlantic Ocean drained Lake Chicago, and the Chicago Lake Plain emerged with essentially its present aspect. Unlike the beaches north of Milwaukee, the beaches of the Chicago Lake Plain were not subsequently tilted by differential uplift of the region.

TILL PLAINS SECTION

The Till Plains section, which covers about four-fifths of Illinois, is by far the largest physiographic division in the state, and in it seven subdivisions are recognized: the Kankakee Plain, Bloomington Ridged Plain, Rock River Hill Country, Green River Lowland, Galesburg Plain, Springfield Plain, and the Mount Vernon Hill Country. The section is characterized by broad till plains, which are uneroded or in a youthful stage of erosion in contrast with the maturely eroded Dissected Till Plains on the older drift-sheets to the west. The outer boundary coincides closely with the margin of the Illinoian drift.

Kankakee Plain.—The Kankakee Plain is a level to gently undulatory plain, with low morainic islands, glacial terraces, torrent bars, and dunes. It is partially fluviolacustrine in origin, but it differs from the lake plains of the Great Lake section in that the lakes which covered it were temporary expansions of glacial floods and did not extensively alter its surface either by deposition or by erosion, except along the courses of strong currents. It could be considered a modified intermorainic basin, floored largely with ground moraine and bedrock.

The district is enclosed by the Iroquois, Manhattan, and Minooka moraines of Cary age on the east and northeast and by the Marseilles and Chatsworth moraines of Tazewell age on the west and south.

Most of the region is poorly drained by shallow low-gradient streams which follow constructional depressions. The two major streams—the Kankakee and the Des Plaines—occupy glacial sluiceways, which, near Kankakee and Joliet, are entrenched in Silurian dolomites. The drift is thick to thin and in the Kankakee region scarcely conceals the bedrock surface.

Bloomington Ridged Plain.—The Bloomington Ridged Plain includes most of the Wisconsin moraines of Tazewell age and is characterized by low, broad morainic ridges with intervening wide stretches of relatively flat or gently undulatory ground moraine. In many places the major moraines rise with gentle slopes, and, although they are conspicuous from a distance, they become less so near at hand, whereas the minor moraines are prominent locally. It was in this district more than in any other that the grass-covered stretches of rolling prairie and extensive swamps, described by early settlers, were most typically and extensively developed (Poggi, 1934, pp. 1–124). The outer boundary of the district follows the outer border of the Shelbyville moraine from the Indiana line westward and northward to the Green River Lowland, beyond which it follows the outer border of the Bloomington moraine to its intersection with Hampshire Ridge of the Great Lake section.

The moraines form a series of loops roughly concentric with the outer boundary of the district and from south to north include the following moraines: Shelbyville, Cerro Gordo, Champaign, Leroy, Bloomington, Normal, Outer Cropsey, Middle Cropsey, Inner Cropsey, Farm Ridge, Chatsworth, Marseilles, and Iroquois. The outer moraines are more widely spaced than the inner moraines, and a series of re-entrants in the south-central part of the district indicate an interlobate relationship of the Lake Michigan and Lake Erie lobes. A temporary lake, Lake Ancona (Willman and Payne, 1942, pp. 213–215), is known to have existed behind the Inner Cropsey moraine, but it did not modify the

ground moraine to any important degree.

The glacial deposits are relatively thick throughout the district and completely conceal the bedrock topography, except locally. Illinoian and older drift are present below the Wisconsin in most places, so that the level aspect of present drift-plains is due largely to the presence of the older drift-sheets, which filled in and covered the irregularities of the bedrock surface (fig. 3).

Drainage development is generally in the initial stage, and most streams follow and are eroding in constructional depressions, many of which cross morainic ridges. Undrained basins are much less numerous than in the Wheaton Morainal Country and occur mainly along the morainic ridges. The valleys of principal streams are larger and more numerous than in the Great Lake section, owing in part to greater areal extent of this division and to somewhat greater age, and they have floodplains bordered by valley-train terraces. The Illinois River, the master-stream of the district, has a broad flat-bottomed valley with steep walls and is bordered by numerous narrow steep-walled valleys with steep gradients. Between the "Big Bend" and Peoria the valley coincides with the large pre-Wisconsin valley of the ancient Mississippi and is wider and has a lower gradient than the upstream part of the valley, which is much younger.

Green River Lowland.—The Green River Lowland is a low, poorly drained plain with prominent sand ridges and dunes. It is bounded on the north and south by the Shelbyville moraine of the Green River lobe and on the east by the abrupt front of the Bloomington moraine (Leighton, 1923, pp. 265–281). Most of the district is modified outwash plain related to the Bloomington moraine, and it is only in the western part of the area that it merges with the Cary valley-train of the Rock River. Some of the sand ridges are in part bars on the outwash plain, but many are true longitudinal dunes with a west-northwest orientation or crescentic parabola dunes. North of Geneseo, remnants of the Shelbyville terminal moraine can be recognized. At the close of the Cary substage the lowland was a great swamp in which the two principal rivers, Rock River and Green River, flowed sluggishly along poorly defined valleys choked with outwash.

The present lowland coincides in large part with a broad bedrock lowland which was occupied by the Mississippi River up to the time of Wisconsin glaciation; and a remnant of the old southern valley-wall forms a prominent bluff on the south side of the present lowland.

Rock River Hill Country.—The Rock River Hill Country is characterized by subdued rolling hill-lands in the stage of late youth to early maturity. It includes the eroded Illinoian drift-plain north of the Shelbyville moraine and Meredosia Valley and a fringe of early Wisconsin drift which lies west of Marengo Ridge.

The Illinoian drift is thin throughout most of the district and is not known to be underlain by older till. Thus the major uplands and valleys are determined primarily by the bedrock surface. The Illinoian drift is without marked ridging, and constructional forms are very localized. In the western part of the district, where it borders the Mississippi Valley, thick deposits of loess and fine sand occur as broad ridges, paha, and dunes on the Illinoian till plain.

The major streams flow radially from a central upland into the Mississippi River on the west and the Rock River on the east and south. Their valleys are relatively broad and steep walled and have

terrace remnants of alluvial fill. The Mississippi River and the upper part of the Rock River occupy large alluviated valleys. Below the mouth of Kishwaukee River, the Rock has cut a post-Illinoian rock gorge, which extends south to the Green River Lowland. Numerous smaller rock gorges are also present along tributaries which locally are superimposed on spurs of the bedrock upland (Hershey, 1893, pp. 314–325). Most of the minor streams are narrow and V-shaped.

Galesburg Plain.—The Galesburg Plain in western Illinois includes the western segment of the Illinoian drift-sheet. The till plain is level to undulatory with a few morainic ridges and is in a late youthful stage of erosion. It is bounded by Meredosia Valley and the Wisconsin drift border on the northeast, by the Illinois Valley on the southeast, and by the Illinoian drift boundary on the southwest. On the northwest a continuation of the district across the Mississippi River into Iowa is implied. Four morainic ridges can be recognized in the district—two occur near the drift border, a third lies near and roughly parallel to the Illinois Valley, and the fourth—the Buffalo Hart moraine—extends northward through the central part of the region. Locally, the Buffalo Hart moraine is a prominent physiographic feature.

The district is drained by streams which flow from a central upland westward into the Mississippi River and eastward and southward into the Illinois River. The larger valleys are steep walled, alluviated, and terraced, except for local narrowing along postglacial gorges. Much of the district is relatively high above baselevel, so that the minor valleys are numerous, deep, and youthful.

The Illinoian drift is generally thick and is underlain by extensive Kansan and Nebraskan deposits, especially along buried preglacial valleys. Most of the irregularities of the preglacial surface were filled in with older drift, so that, in contrast with the Rock River Hill Country, only gross features of the bedrock topography are reflected in the present landscape.

Springfield Plain.—The Springfield Plain includes the level portion of the Illinoian drift-sheet in central and south-central Illinois. It is distinguished mainly by its flatness and by shallow entrenchment of drainage as compared with the more sharply incised valleys of the Galesburg Plain. The southern boundary of the district, which coincides closely with a similar division made by Paul McClintock in 1929 (fig. 1, p. 28), is drawn along a line south of which the drift thins and bedrock topography becomes a controlling factor; the western boundary follows the edge of the Illinoian drift.

Although the greater part of the district is a flat till plain, the morainic features in the western part of the region are much more conspicuous than elsewhere on the Illinoian drift-sheet. They include the Jacksonville and Buffalo Hart moraines and an extensive area of ridged drift in the Kaskaskia drainage basin. The moraines are low and broad, but they are readily recognized because of their continuity and the associated kames and kame terraces. The Kaskaskia ridges lie just to the east of an interlobate area indicated by the moraines and include an irregular assemblage of gravelly ridges and hills with small intervening plains, some of which are old lake basins. A large proportion of the hills and ridges appear to be kames and crevasse-fillings related to stagnant ice conditions (Ball, 1940, pp. 951–970).

Drainage systems are well developed, and the district as a whole is in a late

youthful stage of dissection. The uplands are low with respect to the master-streams, and the valleys are relatively shallow. Most of the principal streams have low gradients and occupy broad alluviated and terraced valleys; the secondary tributaries have wide V-shaped valleys; and the headwaters, flowing essentially on the till plain, have broad shallow valleys and low gradients.

The Illinoian drift is moderately thick and is underlain by older drift except in areas where the bedrock is close to the surface. Only the larger valleys and uplands of the bedrock surface are reflected in the present topography (fig. 3).

Along the southeast side of the Illinois Valley there is a belt of thick loess, with dune-contours characterizing the bluff-margin, but this body of loess thins rapidly to the southeast.

Mount Vernon Hill Country.—The Mount Vernon Hill Country comprises the southern portion of the Illinoian drift-sheet in Illinois and is characterized by mature topography of low relief with restricted upland prairies and broad alluviated valleys along the larger streams. Except for a southern extension of the Jacksonville moraine, glacial land forms are essentially absent. The southern and western boundaries of the district coincide closely with either the outer limits of glaciation or the outer margin of the Carbondale group of the Pennsylvanian system.

A relatively complete drainage system is present, and most streams have broad terraced valleys and low gradients. Natural drainage is good throughout the upland area, but the larger valley bottoms are poorly drained. Extensive aggraded lowlands along the Wabash drainage system to the east and the Big Muddy basin to the west are outstanding physiographic features.

The Illinoian drift in the district is thin, and deposits of underlying older drift are not known to be present except west of Sparta in Randolph County. The present land surface is primarily a bedrock surface of low relief, which is only slightly modified and subdued by a mantle of drift (fig. 3).

Geologic history of the Till Plains section.—Prior to glaciation the Till Plains section had a long and complex erosional history (Horberg, 1946, pp. 179–192). An extensive lowland—the central Illinois peneplain—was eroded in the weak Pennsylvanian rocks of the Illinois basin east of the Illinois River; it was bordered on the west and south by uplands, on which remnants of an older erosion surface are extensively preserved. Just prior to glaciation a system of deep bedrock valleys, many of which are occupied by present streams, were entrenched below the level of the central lowland. The gross features of the section as well as local features in the Rock River Hill Country and in the Mount Vernon Hill Country are determined to a large degree by this preglacial topography. The greater relief and higher elevations in the Rock River Hill Country and in the Galesburg Plain are determined by the preglacial uplands, whereas the low plains in the remaining districts reflect the central Illinois peneplain.

With the advent of glacial conditions and the approach of the Nebraskan glacier, there was probably a change from erosion to aggradation along major streams as the result of increased load and drainage derangements. The oldest deposits along the ancient Mississippi Valley (middle and lower Illinois) and its buried eastern fork, Mahomet Valley, appear to represent this stage. This was followed by the Nebraskan glacial invasion, which is known to have covered at

least a large part of the upland of western Illinois. There is no evidence that the early fills in the preglacial valleys were more than partially removed during the succeeding Aftonian interglacial stage. The Kansan glaciers, which followed, advanced from both the northeast (Labradorean center) and northwest (Keewatin center), probably in that order, and together covered most of the district except for the Rock River and Mount Vernon hill lands. Mahomet Valley and its tributaries in the central part of the section were largely buried and, because of the diversion of drainage, were not reexcavated during the ensuing Yarmouth interglacial stage (fig. 3) (Horberg, 1945, pp. 349–359).

With the advance of the Illinoian glacier from the Labradorean center, the ice attained its maximum extent in Illinois, and the entire Till Plains section was ice-covered. Except in the Rock River and Mount Vernon districts, where older drift is largely absent, the ice moved across a subdued land surface with fills of early drift and during retreat left behind a relatively smooth till plain. Following Sangamon erosion, which was not important except locally, the Wisconsin glaciers of the Tazewell stage covered the northeastern part of the state and formed the glacial landscape which is still so extensively preserved.

DISSECTED TILL PLAINS SECTION

The Dissected Till Plains section in western Illinois is represented by a narrow isolated segment of the Kansan drift-sheet maturely dissected into an upland of high relief. The eastern boundary is determined by the Illinoian drift margin and the southern boundary by an arbitrary line, south of which the drift occurs as patches and is unimportant physiographically. From a regional standpoint, essentially the entire section lies west of the Mississippi River, and the eastern boundary of the section was drawn along the river by Fenneman. The refinement of the boundary here proposed is of only local significance.

The Kansan drift in the area is thin, and the topography closely reflects the ruggedness of the underlying bedrock upland. Valley-train terraces along the Missisippi Valley and thick loess deposits on the adjoining bluffs are features of secondary importance. The geomorphic history closely parallels that of the adjoining Ozark Plateaus and is discussed later.

WISCONSIN DRIFTLESS SECTION

The Wisconsin Driftless section, which constitutes the final subdivision of the Central Lowland province in Illinois, is a submaturely dissected, low plateau bordering the outwash-filled valley of the upper Mississippi River. The eastern boundary follows the edge of the Illinoian drift and is unchanged from Fenneman's classification.

The upland is underlain by the Galena-Platteville dolomite and Maquoketa shale of Ordovician age and by outliers of Silurian dolomite. Flat upland areas, which coincide closely with the top of the Galena-Platteville dolomite, are considered remnants of the Lancaster peneplain.[1] A possible higher surface—the Dodgeville peneplain—may be represented by the crests of mounds and narrow ridges capped by Silurian dolomite. Benches clearly controlled by structure occur along the lower reaches of the principal valleys, where the Maquoketa shale has been stripped from the top of the Galena-Platteville dolomite.

The plateau is maturely to submature-

[1] See Horberg (1946) for bibliography and recent review of the problem.

ly dissected by a number of dendritic drainage systems tributary to the Mississippi. The Mississippi Valley has broad terraced bottom lands and precipitous walls. Most of the minor valleys are youthful, with narrow V-shaped valleys and some with incised meanders. There is considerable underground drainage through small caves and solution channels, but sinkholes and other karst features are not conspicuous. The canyon of Apple River is a prominent local feature resulting from glacial diversion of the former headwaters of Yellow River (Trowbridge and Shaw, 1916, pp. 95–99). As elsewhere, thick loess deposits mantle the bluffs of the Mississippi Valley and thin eastward.

The geomorphic history of the region is largely one of stream erosion and involves numerous uncertainties. The broad outlines, however, may be summarized as follows: (1) development of the Dodgeville surface, probably as a peneplain; (2) rejuvenation and erosional development of the Lancaster surface to a partial peneplain; (3) uplift and entrenchment of the Mississippi bedrock valley and some of its larger tributaries; (4) partial filling of the valley by glacial outwash with a maximum thickness of more than 300 feet; and (5) postglacial erosion.

OZARK PLATEAUS PROVINCE

The Ozark Plateaus province forms a discontinuous upland along the southwest margin of the state and represents the eastern edge of an extensive upland in southern Missouri and northern Arkansas (fig. 2). It includes the driftless and thinly drift-covered cuestas on pre-Pennsylvanian rocks which are structurally and topographically a part of the Ozark dome. Two important modifications of Fenneman's classification are proposed: (1) the Salem Plateau section is expanded northward to include the partially drift-covered Mississippian cuesta in Randolph, Monroe, and St. Clair counties, which is clearly a part of the Ozark dome; (2) the Lincoln Hills section, first distinguished by E. M. Shepard (1907, pp. 8, 10–11) in Missouri and later by W. W. Rubey (1936) in Calhoun County, Illinois, is recognized as a new subdivision. Fenneman included both these areas in the Till Plains section of the Central Lowland province.

Lincoln Hills section.—The Lincoln Hills section includes the partially drift-covered dissected plateau above the junction of the Mississippi and Illinois rivers in western Illinois. It is part of a larger upland which, bisected by the Mississippi, lies partly in Missouri. The principal physiographic feature in Illinois is a maturely dissected central ridge, which forms the watershed between the Mississippi and the Illinois rivers throughout the length of the section. As previously noted, the northern boundary is arbitrary, and the eastern boundary follows the Illinoian drift border. The southern boundary with the Salem Plateau is drawn along the Cap au Grès flexure in southern Calhoun County.

The upland is determined by a subsidiary structure of the Ozark dome, the Lincoln fold, along which the more resistant pre-Pennsylvanian limestones and dolomites crop out. In Illinois the plateau is largely underlain by Osage limestones, of which the Burlington limestone is most important physiographically; and the boundaries coincide quite closely with the Mississippian-Pennsylvanian contact. The southern part of the section is driftless except for loess deposits and a single high-channel filling of outwash gravel, presumably Kansan. It has long been known as the Calhoun

County driftless area. Patchy remnants of Kansan drift are preserved in the northern part of the section.

The plateau surface is rugged and broken by closely spaced valleys and ridges. Restricted areas of flattish to gently rolling upland representing the Calhoun peneplain (Rubey, 1936) are present along the crest of the ridge. The valleys of the Mississippi and Illinois rivers are broad, deeply alluviated, terraced, and have precipitous walls. Most of the minor valleys are narrow, V-shaped, and steeply graded.

Salem Plateau section.—The Salem Plateau comprises the major part of the Ozark dome in southern Missouri, but only two small segments, isolated by the Mississippi River, are present in southwestern Illinois. Both segments are maturely dissected, partially truncated cuestas, dominated by a single central ridge. The northern segment is covered by thin Illinoian drift, but the southern segment lies south of the glacial boundary. In the northern segment an arbitrary boundary with the Shawnee Hills is drawn where the sandstones and conglomerates forming the lower Pennsylvanian escarpment give way to finer sediments and the escarpment dies out, the east margin closely follows the overlapping edge of Pennsylvanian strata, and the northern boundary coincides with the Cap au Grès flexure. The southern segment is delimited from the Shawnee Hills to the east along the contact between Carboniferous and older rocks and follows Fenneman's boundary.

The northern segment is developed on Mississippian strata and lies on the back slope of the Meramec-Osage cuesta, which flanks the Ozark uplift on the north and east. It is underlain by Meramec limestones on the west and north and by Chester strata on the southeast. The plateau is submaturely dissected; and gently rolling summit areas, considered remnants of the Ozark peneplain, occur along the central ridge. Because of the drift mantle, the topography does not appear as rugged as the Lincoln Hills or the Salem Plateau sections. Karst features, developed primarily on the St. Louis limestone, are present at many places within the area. The central ridge forms the watershed for tributary drainage, but the Mississippi and Kaskaskia rivers cross the ridge without regard to structure. The valleys of these two major streams have broad alluvial flats and steep walls, whereas most of the tributaries are youthful.

The south unglaciated segment of the Salem Plateau in Illinois is underlain largely by a thick succession of deeply weathered Devonian chert and cherty limestone formations which on the south are overlapped by Coastal Plain sediments. Structurally, the area is clearly a part of the Ozark dome, but it is complicated by a zone of folds and faults trending north-south and northwest-southeast. A clearly defined physiographic boundary separates the plateau from the Shawnee Hills to the east and north, the contrast being marked by more rugged hills, closer drainage texture, absence of structural control, and higher elevations in the plateau section. Most of the plateau is maturely dissected by intricate dendritic drainage, although small remnants of a flat upland surface representing the Ozark peneplain are preserved throughout the region. The northern part of the segment is drained by streams which head in the Shawnee Hills and flow westward across the plateau into the Mississippi River, whereas in the southern part a central divide separates the Mississippi and Cache Valley drainage. In contrast to other parts of the Ozark

Plateau in Illinois, most of the larger tributary valleys, as well as the Mississippi Valley, are deeply alluviated, and only the secondary tributaries are youthful.

Geomorphic history.—The Ozark Plateaus are essentially a preglacial land surface whose erosional history has continued during the glacial period. The oldest landscape features in the province are isolated summit areas, over 800 feet above sea-level, which may be peneplain remnants correlative with the Dodgeville peneplain of the Wisconsin Driftless section and the Buzzard's Point plain (Salisbury, in Weller, Butts, Currier, and Salisbury, 1920, pp. 47–52) of the Shawnee Hills. An alternative interpretation is that they are simply monadnocks on the lower Ozark peneplain. In either case an extensive surface, called the "Calhoun peneplain" in the Lincoln Hills section and the "Ozark peneplain" in the Salem Plateau section, was developed below the level of these isolated remnants and is responsible for the general accordance of summit levels found throughout the plateau at elevations about 700 feet above sea-level. Near the southern margin of the plateau the Ozark peneplain is believed to transect Wilcox (Eocene) strata and therefore to have been completed sometime during the Tertiary (Flint, 1941, pp. 634–636). The weathering and leaching of the Devonian formations in the southern part of the Salem plateau to depths of about 400 feet is of unusual interest and has been ascribed to prolonged alteration under peneplain conditions (Weller, 1944, pp. 101–102). Following completion of the peneplain, and probably prior to erosion of the Central Illinois peneplain, "Lafayette"-type gravels were spread over its surface. It appears likely that the major preglacial drainage lines were determined at this time and that, with uplift of the peneplain, streams occupying the Mississippi, Illinois, and Kaskaskia valleys became incised without regard to structure. There is no clear evidence of the succeeding central Illinois peneplain and Havana strath cycles in the region, probably because the weaker formations on which they are elsewhere developed are absent. During the glacial period the preglacial topography was modified by alluviation of the major valleys and by deposition of loess on the uplands.

INTERIOR LOW PLATEAUS PROVINCE

Shawnee Hills section.—The Interior Plateaus in southern Illinois are represented by the western part of the Shawnee Hills section[2] and include a complex dissected upland, underlain by Mississippian and Pennsylvanian strata of varied lithology. It is, in the main, the area generally referred to popularly as the "Illinois Ozarks." The northern margin is drawn along a marked topographic boundary which lies along the inner flank of the lower Pennsylvanian (Caseyville) cuesta just within the Illinoian glacial drift boundary, and the southern boundary follows the northern edge of the overlapping Coastal Plain sediments. These are essentially Fenneman's boundaries, the only important modification being a northwestward extension of the section to include the thinly drift-covered Pennsylvanian cuesta in Jackson and Randolph counties.

The section is situated along the southern rim of the Illinois basin, so that the lower Pennsylvanian cuesta com-

[2] The section was originally distinguished by R. F. Flint (1928, pp. 451–457) and named the "Shawnee Hill Section. Fenneman (1938, p. 435) used the name "Shawnee section." The suggested usage of the term "Shawnee Hills section," in the present report is proposed purely for descriptive reasons.

prises the northern part of the region and a dissected plateau underlain largely by Chester (Mississippian) formation comprises the southern part. This regional structure is complicated by faulting and folding, which to a varying degree involved a large part of the area.

The Pennsylvanian cuesta forms a continuous ridge and watershed, extending completely across the state. In most places the ridge is maturely dissected by youthful valleys, but remnants of flat upland are locally preserved on narrow ridge crests throughout the length of the escarpment.

The plateau on Mississippian rocks to the south is maturely dissected, and the larger valleys are alluviated. There are numerous minor escarpments, structural benches, fault-line scarps, and subsequent valleys which reflect local structure and the varied lithology of the bedrock. Only small patches of flat upland are present. Karst features in the St. Louis limestone are present near Cave in Rock, Hardin County, and in south-central Union County.

Geomorphic history.—The erosional history of the region is similar to that of the Ozark Plateaus previously outlined. Remnants of the Ozark peneplain appear to be extensive along the Pennsylvanian escarpment, and local higher summits, especially to the east, may represent an older (Buzzard's Point plain) cycle (Salisbury, in Weller, Butts, Currier and Salisbury, 1920, pp. 47–52). Lower surfaces on the Mississippian rocks, 500–550 feet and 600–650 feet in elevation, are of uncertain origin. Remnants of "Lafayette"-type gravels are found both on the escarpment and at lower elevations south to the Ohio River. A deep weathered zone on the gravels overlain by loess indicates that a long period of stable conditions followed their deposition and that the major period of valley-cutting occurred late in the Tertiary (Weller, 1940, p. 45). Loess deposition and valley alluviation were the principal events during the glacial period.

COASTAL PLAIN PROVINCE

The Coastal Plain in Illinois includes the southern tip of the state and is underlain by unconsolidated Cretaceous and Tertiary sediments, which overlap the older Paleozoic rocks to the north. Three physiographic subdivisions are recognized: (1) and (2) the coextensive alluvial plain of the Cache and Mississippi valleys and (3) the Cretaceous hills between the Cache Valley and the Ohio River. The alluvial plains are characterized by terraces and recent floodplain features. The Cretaceous hills are maturely eroded into a low upland of gently sloping knolls and ridges. Outwash and alluvium extend far up tributary valleys, so that the upland is partially buried and certain segments are essentially isolated.

The earliest events in the geomorphic history of the region are indicated by remnants of "Lafayette"-type gravels which occur in the Cretaceous hills. The erosion surface at their base is evidence of a long period of denudation, during which the Coastal Plain deposits were lowered and stripped back. This was followed by deposition of the gravels, their weathering under stable conditions, establishment of major drainage lines, and final dissection of the bedrock topography. Prior to glaciation, Cache Valley was occupied by the Ohio River and the present Ohio Valley was occupied by the Cumberland and Tennessee rivers. During Illinoian or possibly Wisconsin time the valleys were aggraded to the level of the divide between the Ohio and the Cumberland rivers at Bay City in south-

ern Pope County, and the present lower course of the Ohio was opened. Both courses were kept open during subsequent stages, so that flood waters still pass through Cache Valley, and it was only in relatively recent time that the southern channel became the permanent course of the river.

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THE FORMATION OF BEACH CUSPS

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ABSTRACT

In former attempts to explain the development of beach cusps, stress has been laid on the erosion of the beach. It is argued in this paper that concomitant deposition on the horns is at least equally important. Refraction of the swash in building the cusps in emphasized, and an attempt is made to explain the rhythmic, roughly equidimensional nature of cusps.

INTRODUCTION

Beaches sometimes show regularly spaced, crescentic accumulations of materials, ranging from sand to cobbles. These formations are now generally termed "beach cusps." The projecting parts are more or less triangular, with a rounded apex extending into the water and with curved bays between the promontories. The basal, or landward, parts of the triangle may be horizontal, with the apex sloping down the beach. In many cases the cusps are of coarser material than that of the remainder of the beach, and they either merge gradually into the latter or are set off sharply from them. The height is usually very slight but may also attain more than a meter, and the distance between cusps ranges from a few centimeters to several dozens of meters. The apex may protrude a few centimeters to several meters, and the relative depths of the bays are also variable. In front of the bays the foreshore is built out under water in a delta shape (fig. 1).

Although cusps may develop on many types of beach, they are most common on slightly concave stretches of the coast. Observation has shown that they may be formed or destroyed in a few hours, small ones even more quickly. Each change in size of the waves results in the development of a new series of cusps to fit the altered conditions. According to Douglas Johnson (1919), the interspace is roughly doubled for a doubling of the wave height.

O. F. Evans (1938) made observations on lake shores and divided the observed cusps into five classes. Two of these occur as individual accumulations, not as a rhythmic series. His "giant cusps" are formed by erosion and deposition during storms. Very small cusps a few inches apart are the result of a gentle onshore "slop" of a "dead sea" and are really a series of rill marks. None of these will be considered in this paper, but the giant cusps are practically the same as his fifth class. These "ideal cusps," which usually occur in series, are similar to those on sea beaches. Evans showed by measurement that the interspace generally varies about 50 per cent, occasionally over 100 per cent. Although these variations are appreciable, they confirm the conclusion that the phenomenon is essentially of a rhythmic nature. Any attempt at explanation must show that the process is not arbitrary in its action but that it contains elements regulating the cusp-interspace.

ORIGIN OF BEACH CUSPS

The origin of beach cusps is not fully understood. Escher (1937) was able to produce cusps experimentally by the action of ordinary waves, combined with standing waves, the latter at right angles

to the model beach. This explanation cannot be applied to beach cusps until proof is given of the occurrence in nature of strongly developed standing waves with wave lengths comparable to the size of cusps. No records of such standing waves are known to the present author, and it appears highly doubtful whether they could develop over distances of hundreds of meters. According to Johnson were not true cusps, but of the nature of his "very small cusps."

This explanation is certainly more satisfactory than many other suggestions that have been offered. But several points remain obscure. In the first place, the beach as a whole is not eroded during the development of cusps, but in Johnson's theory only erosion is mentioned. Evidently, accumulation goes on concomi-

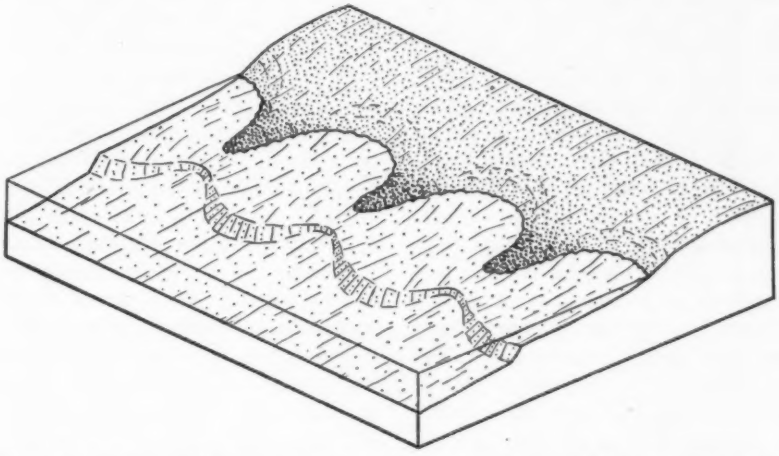

FIG. 1.—Block diagram of beach cusps, showing the horns (in this case consisting of coarse material) and the submarine deltas corresponding to the embayments. (Mainly after Timmermans.)

(1919), erosion of the beach by the swash plays an important part. Slight irregular depressions undergo erosion by the waves because more water rushes in and out than up and down the adjoining even slopes. The larger a depression becomes, the more strongly it will grow. These depressions are enlarged until they together occupy the whole beach, with narrow tongues between. The coarse material tends to be thrown up the beach and onto these projecting cusps. Johnson (1919) and Timmermans (1935) succeeded in producing small cusps experimentally in a tank with a wave-producing apparatus. According to Evans, these tantly with erosion, and the horns must represent not merely buttresses left standing by the erosive action in the bays but prograded areas where most of the eroded material comes to rest. The fact that the cusps generally consist of material differing from the remainder of the beach is further evidence of the selective transportation from bay to horn and the outbuilding of the latter.

In the second place, it is not evident why a stable condition is attained, once the cusps fitting the waves are developed. Johnson says (1919, p. 483): ". . . . enlargement will continue only so long as the impulse toward growth im-

posed on the more favored channels is sufficiently great to overcome the tendency of their neighbors to enlarge. Equilibrium will be established when adjacent channels are approximately the same size, and at the same time of a size appropriate to the volumes of water traversing them." But, as further erosion would result in a larger volume of water running in and out of an embayment, one may ask what is the meaning of the size being "appropriate to the volume of water." Neither is it clear why the growth of its neighbors should impede the enlargement of a bay.

As soon as it is admitted, however, that the horns are built out by materials washed from the bays, it is obvious that there must be a close relation between the development of adjoining bays. Wherever one is enlarged, the concomitant growth of its horns must tend to obliterate its neighbors. In the sentence quoted from Johnson, we should make a small alteration as follows: "Enlargement of the bays *and cusps* will continue only so long as the impulse towards growth imposed on the more favored channels is greater than with its neighbors." It then remains only to be shown that, starting from small irregularities, the larger ones are, first, at an advantage but that, with increasing size, this impulse toward growth must eventually disappear, so that the less developed bays will then tend to enlarge at the cost of the larger, overgrown channels. Only in this manner can an evenly spaced series of cusps be accounted for.

Before attempting to show that this relation actually exists, the process by which the horns are prograded must first be found. The present writer has observed that *refraction of the swash*, as it sweeps into the embayments and fans out onto the sides of the protruding horns, is also an essential element in the production of the cusps. This causes pebbles to be washed sidewise out of the troughs onto the horns.

The backwash is somewhat concentrated toward the horn by the same process. The sidewise component of the uprush is directed toward the cusp from both sides (pl. 1, *A*); and, because we are dealing with swash and not with oscillation waves, an actual piling-up of water over the area of the horn is brought about. When the water begins to flow back under the influence of gravity (*a*–*f* in fig. 2), it must trend somewhat in the direction perpendicular to the shore line (*a′*–*f′*) in consequence of the accumulation of water over the area *DEF*.

Hence a particle of water will follow a curved course over the cusp, somewhat similar to the movement during normal beach drifting. This curved path, directed toward the apex of the cusp, carries the pebbles out toward the tip of the horn. But, because waves tend to throw pebbles high up onto the beach and carry the sand fraction away, the pebbles are deposited on the cusps, and the sand is rolled back into the bay. The next rush of water will bring part of the sand onto the beach of the embayment. The remainder will be dropped on the delta scallops in front of the bays.

Another consequence of the refraction of the swash is that the strengthened backwash over the cusps impedes the next swash on the horn and thus helps to concentrate erosion in the embayment. A complication arises because the swash is impeded by the backwash more strongly in the bays and on the horns alternately. The writer has not been able to make sufficient observations under varying conditions to ascertain whether or not this alternate action is the rule and forms an essential process in the develop-

PLATE 1

A

A. The same as Pl. 1, *B*, at advanced stage in uprush of the swash. Note the marked refraction of the swash and advance of the water from both sides on the cusps.

B

B, Beach cusps, Alum Bay, Isle of Wight. Apex to apex is 8 meters.

PLATE 2

A

A, Waves playing on cusps in Sandown Bay, Isle of Wight, at low tide. Concentration of backwash at intervals corresponding to the rhythm of the intercusp space.

B

B, The same as *A* at slightly later stage in the uprush of the swash.

ment of cusps. The photograph (pl. 2, *A*) was made during low tide, when the waves were playing only at the lower end of the cusps. It shows the concentration of the backwash at intervals corresponding to the rhythm of the intercusp space. Plate 2, *B*, shows a slightly later stage in the uprush of the swash. appears to exist, but they may be observed in two sets, one directly below the other, where there is no space for a ridge in between (see Johnson, fig. 142). Moreover, Gellert (1937) saw cusps forming during a storm on the German North Sea coast, where no ridge existed. As I have made no observations on lake

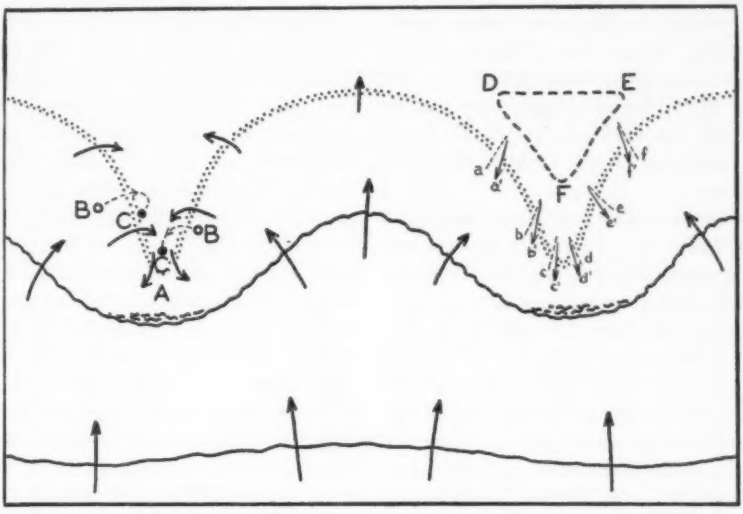

FIG. 2.—Diagram showing the diffraction of the swash. Pebbles are rolled from *B* to *C*. At *A* the powerful backwash of the foregoing wave impedes the swash and increases the diffraction. *a*–*f* = slope, *a'*–*f'* = direction of backwash owing to the piling-up of water over the area *DEF*.

Although Evans (1938) speaks of the "parabolic path" of the water and refraction around the cusp apex, he does not show in what manner this influences the development of cusps.

Evans came to the same conclusion on the origin of cusps as did Jefferson, who also studied lake-beach cusps. Both authors were able to show that on lake shores the breaching of a beach ridge or the edge of a low-cut bank is essential to the formation of cusps. On sea shores with tidal movements this is apparently not a necessary condition. Not only are cusps frequently found where no ridge shores, I will deal only with the case of shores with tides.

Here a certain amount of erosion is probably also essential to the formation of cusps but not to the breaching of a ridge. Thus Timmermans (1935) found that on the Dutch coast and in his experiments cusps developed only where the beach was steeper than normal and erosion took place.

SPACING OF CUSPS

We must now return to the question of even spacing. It is obvious that the degree of refraction and also its conse-

quences are closely related to the size of the waves and the slope of the beach. A small cusp cannot influence the shape of a powerful wave, and the horns will be attacked and washed away. Small waves breaking on large cusps are spaced too closely together so that there is no time for the backwash of the first to run out of the embayment before the next one comes rolling in. Moreover, they carry insufficient water to influence large cusps. Smaller irregularities will then be picked out and developed separately into a new set of narrowly spaced cusps, better adjusted to the small volume of the waves.

This rather vague relation between the size of waves and cusps can be worked out in greater detail. In the first place, it is evident that a certain size of wave cannot erode the bays beyond a certain depth of water because the volume of water passing in and out is constant, and, consequently, with increasing depth the current becomes too sluggish to move the sediment. The distance to which a bay can be eroded into the beach is thereby also limited, because, with a limit set to the depth, the outward slope of the bottom becomes so slight that gravity is unable to cause sufficient current for dragging sediment outward. In the second place, broadening of the bay delivers more material for building out the cusps. This, in turn, must cause prograding of the embayment floor and consequent weakening of the erosion in the bay. In the third place, outward growth of the horns must result in stronger attack by the swash, because the protecting influence of the backwash will decrease. This means that the material carried out toward the apex is brought back into the channel. It then moves back up the bay to the beach.

Summarizing the deductions given above leads us to the following explanation of cusp formation. On a smooth beach a regular train of waves will cause a succession of swash flows. These flows encounter slight depressions and start to erode them, while the backwash carries some sediment out of the embayments to build deltas opposite them. As long as the depth of water in a bay is so small that the water passing in and out is able to carry the sediment along, the enlargement will continue both horizontally and vertically. As the channel becomes deeper and its side slopes steeper, the breadth also tends to increase. But, when the depth in the outer portion of the bay approaches a certain limit, erosion gradually slackens. Refraction of the swash results in transport of material toward the sides and causes prograding at these points. The coarse material tends to be pushed back up the beach of the embayment and out along the developing cusps. Growth of the bays and prograding of the cusps must gradually decrease when the maximum depth in the central area of the bay has been attained.

In the meantime, adjacent channels have been subject to the same process. Where these are so close together that the two natural spheres of growth overlap, a rivalry develops. This should tend to push the bays farther apart, because the material eroded from the one and dumped onto the intervening cusp must tend to encumber the growth of the other. Erosion in the latter will be shifted slightly to the opposite side and thus cause the entire bay to move away.

Where two neighboring bays are farther apart, their adjacent cusps will not coalesce and will leave a small space between. In this space refraction begins, and a new bay will start to form.

As long as the maximum depth of water in a bay has not been attained, the

tendency to enlarge will be greater than in an adjacent, shallower indenture. The larger one will grow at the expense of the smaller one. This relation is reversed when the depth approaches its maximum value. Then a smaller bay is more powerfully eroded and tends to encroach on its overgrown neighbors. In this manner a balance will be reached when all bays have attained the maximum depth for the size of waves playing on them. They will push each other aside or obliterate some until this stage is attained and a regular rhythmic pattern has been evolved (fig. 3, pl. 1*B*).

The writer believes the process here pictured accounts for the main phenomena of cuspate beaches: the regularity of the pattern, the growth in size with increasing wave dimensions, and the accumulation of coarse material in the horns. But, although a step forward may have been made by this working hypothesis in explaining these fascinating features of the shore, several problems still await solution. The differences in size and shape, slope and grain size, are evidently interrelated, but in what manner is not known.

Neither has it been made clear why

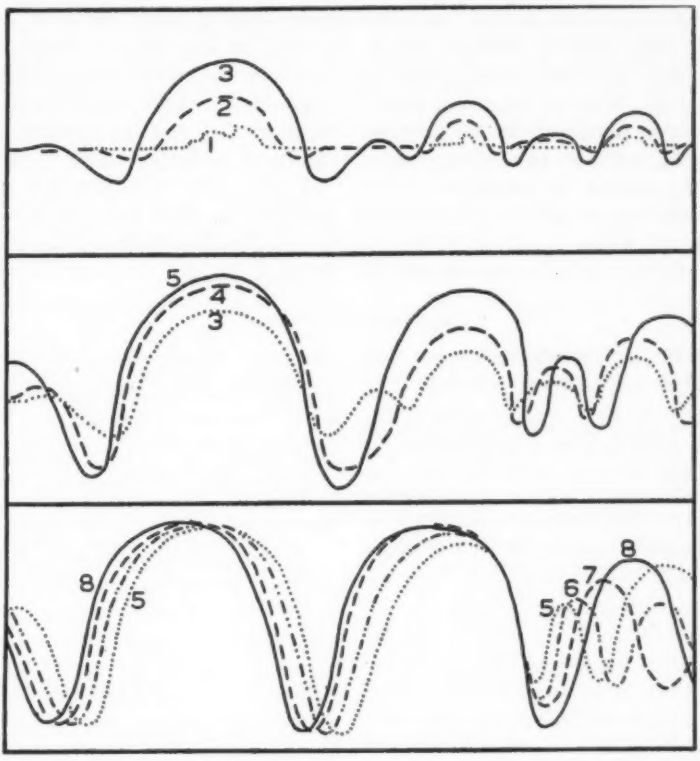

FIG. 3.—Successive stages in the development of cusps from irregular indentures of the beach to an almost regular, rhythmic pattern.

cusps are of only exceptional occurrence. Johnson (1919) and Timmermans (1935) both believed that waves directed straight onshore are most effective in building cusps. Thus the latter found cusps on the Dutch coast only during periods of offshore winds when regular small waves advance straight toward the shore. This may explain why bays are favorable sites, because oblique waves cannot form so easily (Timmermans, 1935, p. 364). But, on the other hand, onshore waves may play for a long time on a beach without producing cusps. Moreover, several investigators have shown that cusps may be formed by oblique winds also.

The influence of tidal movements is another aspect in need of further study. Timmermans (1935) believed that the cusps are first formed during rising tide and then lengthened seaward during the falling of the water. F. P. Shepard maintains (personal communication) that the tides are important. Their action is evidently not necessary, because it has already been pointed out that cusps are formed on lake shores, where they are related to the breaching of a ridge. A special case in which erosion may be the main cause is that of boulder cusps, as described by Butler (1937, pp. 442–453).

Finally, there is an almost total lack of quantitative data and the flow of the water should be studied in much greater detail.

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THE DISTRIBUTION OF OXYGEN IN THE LITHOSPHERE

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ABSTRACT

Oxygen, which makes up more than 90 per cent by volume of the total lithosphere, shows the highest concentration in the outer shell. The regular decrease with depth represents an approximation to thermodynamic equilibrium. When highly oxidized surface rocks are brought down to great depths, oxygen will be squeezed out of the mineral lattices and returned to the surface. Therefore, the deeper parts of our globe cannot become oxidized.

THE VOLUME RELATIONS OF OXYGEN IN THE LITHOSPHERE

The building bricks of the crystals are atoms (or ions); and in the crystalline edifice, as in other constructions, the largest bricks are the most important. In the late twenties it was recognized from X-ray studies that oxygen is among the very largest constituent ions of the rock-forming minerals; and it was appreciated that the oxygen ions, which had been generally neglected by chemists, were of fundamental importance in the physical chemistry of minerals.

The dominant role of oxygen in the lithosphere is demonstrated in table 1.

The igneous rocks and the whole lithosphere are to be regarded as essentially a packing of oxygen ions. The accumulation of this huge volume of oxygen is made possible through the cations, which occupy the interstices and, with their electrical charges, keep the whole structure together, although their volume is comparatively insignificant.

It is interesting to note that the lithosphere contains more oxygen (in terms of atom, weight, and volume percentages) than does the atmosphere. Thus the lithosphere appropriately might be called the "oxysphere" (Goldschmidt, 1928).

Estimates of the chemical composition of the interior of the earth are uncertain—indeed, they are based on extrapolations into the unknown. According to our best information, the outer shell, the so-called "crust," is made up of the following concentric layers which inperceptibly merge into one another:

1. Sediments, which are partly penetrated by
2. "Sial," which extends to a depth of approximately 25 km. Its composition corresponds to the computed average for all igneous rocks, with granitic rocks at the top and gradually changing downward into
3. "Sialma," which occupies the depth-range 20–70 km. and whose average composition is that of a plateau basalt.
4. At still greater depth, heavy basic silicates prevail. This region has been referred to as the "eclogite or peridotite shell." Typically, olivine is present.

Average figures for the oxygen content of the several shells in the earth are listed in table 2. The results are shown graphically in figure 1.

CONDITIONS OF THERMODYNAMIC EQUILIBRIUM IN A FIELD OF GRAVITATION

I shall endeavor to show that the small but consistent drop in oxygen with depth as demonstrated by these data represents an approximation to thermodynamic equilibrium.

In the chemical laboratories we are wont to consider equal vapor tensions in all phases as necessary and sufficient for equilibrium. In a gravitational field this is different, a fact which has been pointed out and emphasized by Ramberg (1944*a*; 1944*b*, pp. 98–111; 1945, pp. 307–326; 1946, pp. 13–29). Various objections, mostly due to misunderstandings, have been raised to Ramberg's principle. To clarify the issue, the following restatement of the principle is offered.

TABLE 1

CHEMICAL COMPOSITION OF TYPICAL ROCKS, EXPRESSED IN VOLUME PERCENTAGES

	Average Basalt*	Average Igneous Rock†	"Ichor" Granite‡	Ionic Radii
O	91.11	91.83	92.12	1.32
Si	0.70	0.83	0.92	0.39
Ti	0.12	0.05	0.02	0.64
Al	0.74	0.79	0.76	0.57
Fe	1.47	0.58	0.21	0.67 0.82
Mg	1.09	0.58	0.09	0.78
Ca	2.78	1.50	0.45	1.06
Na	1.28	1.64	1.75	0.98
K	0.70	2.19	3.68	1.33
Total	99.99	99.99	100.00	

* Average of Plateau basalt, 50 flows. Recalculated from Daly (1927).

† Recalculated from Clarke and Washington (1924).

‡ Typical pre-Cambrian "petroblastic" granite from Birkeland, southern Norway (Barth, 1948).

All three analyses are recalculated *on a water-free base.* Manganese is included in iron.

TABLE 2

OXYGEN IN PERCENTAGE BY VOLUME

Air	20.95
Ocean	96.87
(Quartz	98.73)
"Ichor" granite	92.12
Average igneous rocks	91.83
Average basalt	91.11
(Olivine*	90.00)

* 90 Mg_2SiO_4, 10 Fe_2SiO_4.

We will consider a vertical section through a homogeneous part of the lithosphere: The chemical potential of a species depends here on its position in the vertical section. The differential equation relating the chemical potential, μ, of a chemical species with its position in the gravitational field is

$$\left(\frac{\partial \mu}{\partial x}\right)_{P,\ T} = M\ ,$$

in which M is the molecular weight of the species and x is the vertical distance.

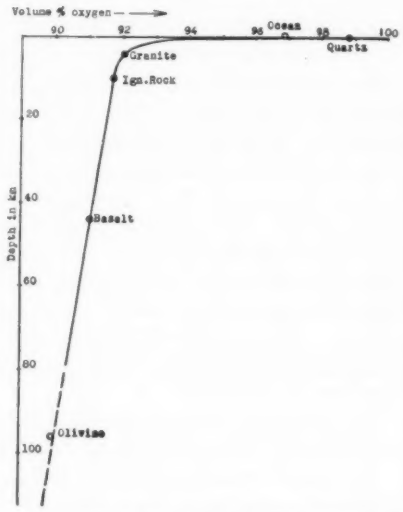

FIG. 1.—The volume percentages of oxygen at various depths.

Consequently, at the depth, x, the chemical potential is

$$\mu_x = \mu_0 - M \cdot X\ . \qquad (1)$$

Thus, the lower the position, the lower the chemical potential. In order to restore equilibrium, the chemical potential must be raised correspondingly. The raise may be effected by the superincumbent load, P:

$$\mu_x = \mu_0 + PV\ , \qquad (2)$$

in which V is the fictive volume of the species. Obviously, the term $M \cdot X$ in

equation (1) corresponds to the loss of potential mechanical energy per mol over a distance X; but it also corresponds to the weight of the superincumbent load if the chemical species with which we are dealing is overlain by itself.

The oxygen ions in the mineral lattices of the lower part of the lithosphere show a chemical potential which is higher than that required to restore equilibrium; for the overlying rocks exhibit a much higher average molecular weight than that of oxygen; this means lack of equilibrium. A manifestation of the disturbed equilibrium and of the high μ-values thus generated at great depths is a high vapor tension and a consequent tendency of the oxygen ions to escape and migrate upward to places of lower chemical potential.

Stated in simple language, it means that, at high pressures, oxygen is squeezed out of the crystalline lattices, whereby compounds containing less oxygen are formed. Thus is established a vertical oxygen gradient in the lithosphere.

Ramberg (1946) has pointed out that water-bearing minerals are not stable at greater depths but give off water, which migrates upward. Nor are high oxides of iron stable; at greater depths they dissociate, with formation of oxygen, which also migrates upward.

THE DEPTH RELATIONS OF Fe_2O_3, Fe_3O_4 FeO, AND IRON

As an example of this process, we can calculate the depth at which oxygen is squeezed out of the iron oxides.

Let us start with a vertical section of the lithosphere containing pure hematite, Fe_2O_3. For the sake of simplicity in this theoretical discussion we shall disregard the upper zone of weathering and oxidation; we postulate an ideal surface of demarcation separating this zone from deeper-lying material that is not in physical contact with, or exposed to, the oxygen of the surface waters and of the atmosphere. Likewise, for the sake of simplicity, we assume that the whole section is of uniform temperature.

The chemical potentials of the oxygen ions in the hematite lattice increase with the superincumbent load in accordance with equation (2). Obviously, we have

$$P = X \cdot d_h$$

and

$$V = \frac{M}{d_0} \text{ (by definition)},$$

where d_h and d_0 are the specific gravities of hematite and oxygen respectively (see table 4).

Substituting in equation (2), we obtain as an expression of the actual increase of the chemical potential of oxygen with depth the following equation:

$$\mu_x = \mu_0 + M \cdot X \cdot \frac{d_h}{d_0}. \qquad (3)$$

This expression should be compared with equation (1). It is seen that the factor d_h/d_0 is a measure of the excess load acting on the oxygen ions. Since $(d_h/d_0) > 1$, it follows that μ_x has a higher value than that corresponding to equilibrium conditions; the crystalline lattice reacts on this by expelling some of the oxygen ions, until the lattice is so poor in oxygen as to become crystallographically unstable; then it inverts into a Fe_3O_4-lattice,[1] or into a FeO-lattice, and eventually into metallic iron.

The task before us is, therefore, to evaluate numerically the increase with depth of the chemical potential of oxygen in layers, or earth shells, composed of Fe_2O_3, Fe_3O_4, and eventually FeO.

[1] The possible effect of the "reversed" specific-gravity relations of hematite and magnetite will not be discussed here.

The chemical potential at the surface has the following general relation to the vapor tension, p:

$$\mu = RT \cdot ln\, p + \text{constant} ;$$

in all numerical calculations we shall use log p as a measure for μ. All the necessary constants are compiled in tables 1, 3, and 4.

The general form of the equation is:

$$\log p_x = \log p_0 + X \frac{M}{RT} \left(\frac{d_1}{d_2}\right) \log e . \quad (4)$$

At 500° C. we find the relation between depth, X, and the vapor tension, p_x, of oxygen in:

A layer of hematite: $\log p_x = X \cdot 6.76 \cdot 10^{-7} - 28$;

A layer of magnetite: $\log p_x = X \cdot 7.05 \cdot 10^{-7} - 30$.

TABLE 3*

VAPOR TENSION EXPRESSED IN ATMOSPHERES†

	500° C.	560° C.	728° C.	1150° C.
Oxygen in Fe_2O_3 (p_0)	10^{-28}	10^{-24}	10^{-16}	$10^{-4.2}$
Oxygen in Fe_3O_4 (p_0')	10^{-30}	$10^{-26.7}$	10^{-20}	10^{-10}
Oxygen in FeO (p_0'')	Unstable	$10^{-26.7}$	$10^{-20.7}$	$10^{-12.6}$

* In this table the densities of oxygen have been computed from the ionic radii and the densities of the several iron oxides. In the following transformations the molecular volumes are shown:

$$\begin{array}{l} 3\ Fe_2O_3 \rightarrow 2\ Fe_3O_4 + O \\ \underbrace{3 \times 30.4}_{91} \rightarrow \underbrace{2 \times 44.5}_{89} \end{array}$$

$$\begin{array}{l} Fe_3O_4 \rightarrow 3\ FeO + O \\ 44.5 \rightarrow 35.9 \end{array}$$

These equations indicate that, in the transformation hematite → magnetite and magnetite → wüstite, the removal of one oxygen ion corresponds to a decrease in volume of 2 units and 8.6 units, respectively. The two values are quite different, and only the last value compares with the data of table 4, from which the fictive volume of oxygen in magnetite is computed as

$$\frac{\text{Atom weight}}{d_o'} = \frac{16}{1.62} = 9.9 .$$

Thus, for the purpose of the present paper, which is mainly a discussion of the transformation, magnetite → wüstite, we can, without serious error, use the constants of table 4.

† Computed and extrapolated from Internat. Critical Tables, from Landoldt-Börnstein's Physicalische Tabellen, and from recent papers by: L. S. Darken and R. W. Gurry: The system iron-oxygen I and II: Jour. Am. Chem. Soc., vol. 67, p. 1398, 1945; vol. 68, p. 798, 1946.

TABLE 4

SPECIFIC GRAVITIES

Hematite (d_h)	5.26	Oxygen in hematite (d_o)	1.71	$d_h/d_o = 3.07$
Magnetite (d_m)	5.18	Oxygen in magnetite (d_o')	1.62	$d_m/d_o' = 3.20$
Wüstite (d_w)	6.00	Oxygen in wustite (d_o'')	1.65	$d_w/d_o'' = 3\ 64$

Molecular weight, M, of oxygen = 0.032 kg · mol^{-1};
Gas constant R = 82 kg. · cm · mol^{-1} · degree^{-1}.

By using the above constants, the depth, X, is found in centimeters.

The condition of equilibrium between a surface layer of hematite and oxygen ions at a depth, X, is

$$\log p_x = X \cdot 2.20 \cdot 10^{-7} - 28 .$$

These three curves are shown graphically in figure 2.
At 728° C. the corresponding equations are:

In a layer of hematite: $\log p_x = X \cdot 5.22 \cdot 10^{-7} - 16$;

In a layer of magnetite: $\log p_x = X \cdot 5.45 \cdot 10^{-7} - 20$;

In a layer of wüstite: $\log p_x = X \cdot 6.20 \cdot 10^{-7} - 20.7$;

The equilibrium condition: $\log p_x = X \cdot 1.70 \cdot 10^{-7} - 16$.

At 1150° C. the corresponding equations are:

In a layer of hematite: $p_x = X \cdot 3.7 \cdot 10^{-7} - 4.2$;

In a layer of magnetite: $p_x = X \cdot 3.84 \cdot 10^{-7} - 10$;

In a layer of wüstite: $p_x = X \cdot 4.3 \cdot 10^{-7} - 12.6$;

The equilibrium condition: $p_x = X \cdot 1.2 \cdot 10^{-7} - 4.2$.

The results of these computations demonstrate, theoretically, that the oxygen ions are in equilibrium with the hematite lattice only in the uppermost layer (disregarding again the zone of sedimentation and oxidation). As soon as the pressure becomes appreciable, magnetite will form.

The magnetite layer in a crust composed only of iron oxides will reach down to 41 km. at 500° C., to 107 km. at 728°, and to 220 km. at 1150°.

Now we shall leave the hypothetical iron oxide crust and discuss the conditions in an idealized lithosphere.

We assume that hematite is unstable at any depth below the zone of sedimentation and oxidation.

Magnetite is stable to greater depths. The specific gravity of the lithosphere at greater depths is approximately $d = 3.2$. Consequently, the ratio: $d/d_o = 2.0$, and the calculated depth at which magnetite is in equilibrium with a top layer whose vapor tension is like that of hematite is:

At 500° C., 91 km.,

At 728° C., 235 km.,

At 1150° C., 483 km.

At still greater depths, magnetite is unstable, and, if the temperature at these depths is as low as 500° C., then metallic iron will become stable at a depth exceeding 91 km. At temperatures above *ca.* 600° C. the magnetite zone will be underlain by a layer in which wüstite is stable, before one reaches the region of stable metallic iron.

THE APPLICABILITY OF THE CALCULATIONS

The calculations of the various depth ranges are highly theoretical, for they are based on assumptions that correspond to

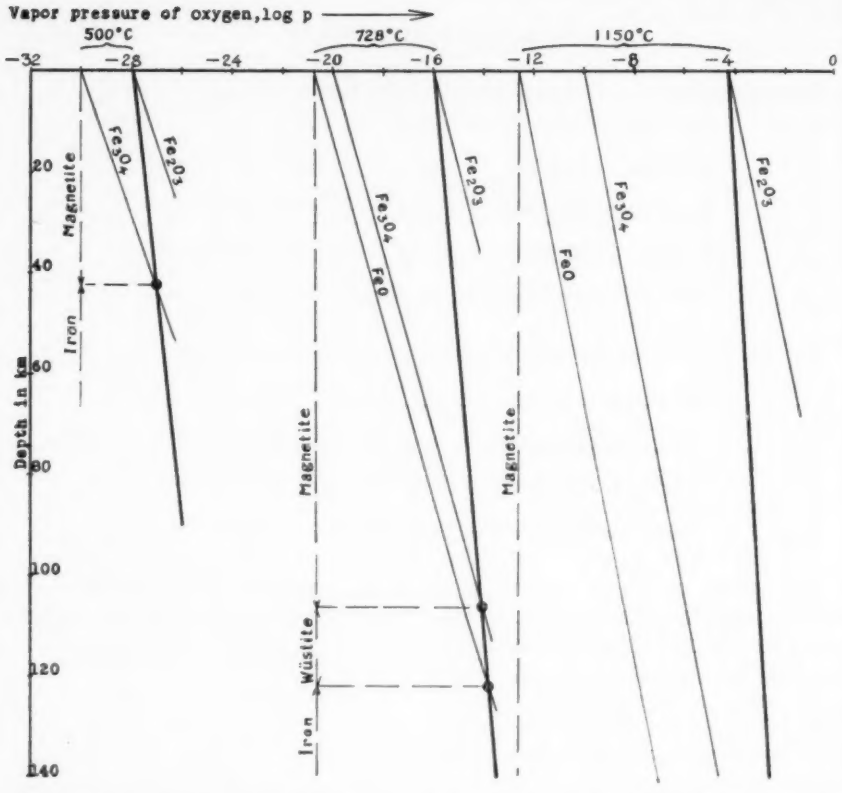

FIG. 2.—Graphical representation of the increase of oxygen pressure with depth, at three different temperatures, in hematite, magnetite, and wüstite. For each temperature is also given, in heavy lines, the curve representing the increase of oxygen pressure corresponding to equilibrium with a top layer of hematite. As seen from these curves, the oxygen tension in hematite increases with depth much faster than that which corresponds to the equilibrium tension. Therefore, hematite is theoretically unstable at any depth and inverts to (oxygen-rich) magnetite.

At 500° C. magnetite is stable to a depth of 41 km. At greater depth the oxygen tension in magnetite becomes too large, the crystalline lattice breaks up, and metallic iron becomes stable. (Wüstite, or FeO, is unstable at this temperature.)

At 728° C. the magnetite layer extends to a depth of 107 km. It is followed by a thin layer of wüstite in the range 107–122 km. At greater depth metallic iron is stable.

At 1150° C. the corresponding demarcation surfaces are at 220 km. and 305 km., respectively. These numerical values are correct only for a hypothetical crust composed of pure iron oxides.

idealized conditions never met with in nature. It is important, therefore, to discuss how far these calculations apply to natural conditions.

Obviously, many chemical elements bear no relation to the law of distribution implied in equation (1). Uranium, for instance, should be found in great depths only, but the opposite is the case. Uranium is a typical "lithophile" element, i.e., the chemical affinity relations of uranium bring it into granitic rocks rather than into the deeper strata, in spite of the gravity relations.

Now oxygen represents a different problem. Because of its great bulk in the lithosphere (= oxysphere) it can be regarded as a "solvent" capable of dissolving the various cations partly in the crystalline state but, at greater depths, in the liquid or glassy state.

Broadly, the "solute" (= all cations taken as an average) shows a tendency to distribution in accordance with equation (3). Thus, if one started with a homogeneous distribution, the tendency would be for the heavier cations to move downward, entailing a relative concentration of oxygen at the top. We have previously considered migration of oxygen upward; this migration is only relative. Owing to the great bulk of the oxygen ions, they probably furnish a rather stationary medium through which the smaller cations move. Through the migration of the cations a vertical oxygen gradient is established, and thermodynamic equilibrium is attained.

Instead of presenting the vertical gradient in terms of volume percentages of oxygen, as was done in figure 1, one might, with advantage, illustrate the same fact by comparing the total number of cations contained in equal rock volumes from various depth.

The data of the present paper make it clear that the volume relations of the ordinary rock types are almost wholly determined by the oxygen ions. If one wants to compare rocks of equal volumes, one can compare rocks containing the same amount of oxygen ions. As a standard comparison unit it is expedient to choose a volume comprising 160 oxygens, for in an ordinary rock nearly 100 cations are associated with 160 oxygens. We shall call a rock unit containing exactly 160 oxygen ions the "standard cell."

TABLE 5*

NUMBER OF CATIONS IN A CELL MADE UP OF 160 OXYGEN IONS

	Number of Cations	Number of Oxygen Ions
Quartz	80.0	160
"Ichor" granite	96.2	160
Average igneous rocks	99.3	160
Average basalt	102.4	160
Olivine	120.0	160

* The average rock analyses used in tables 1 and 2 are used here. The water content has been neglected, to eliminate the effect of the hydrogen ion.

A comparison of the standard cells of rocks of the various depth zones is afforded by table 5, which convincingly demonstrates the increase in cations with depth and thus, in different terms, illustrates the same thing as figure 1.

We conclude, therefore, that the distribution of oxygen in the lithosphere is roughly in accordance with equation (3) and that the mechanism of establishing the equilibrium is a migration of cations in a (glassy) solvent of relatively stationary oxygen ions.

Likewise, the distribution of the iron ions may be expected to be governed roughly by equation (3). Although many additional factors most certainly will bring about great local and regional deviations, it is a fact that iron is an element showing, at the same time, lithophile,

chalcophile, and siderophile tendencies and will therefore behave more like an "average" cation than many of the other elements.

THE INORGANIC CYCLE OF OXYGEN

The discharge of oxygen from the deeper parts of the earth's crust is a geochemical process of great importance. Goldschmidt (1938) has regarded water, CO_2, and large quantities of Cl, F, B, S, and Se as products of degassing of the lithosphere. At one time oxygen also was formed through degassing.

Pertinent data have been published by Tammann (1924, p. 17), who demonstrated rather convincingly how great quantities of free oxygen must have developed immediately after the formation of a solid crust. At high temperatures and with a high partial pressure of steam in the atmosphere, the dissociation of steam was at that time great enough to provide for all the free oxygen now found in the atmosphere (Wildt, 1942, pp. 151–159).

Goldschmidt has criticized Tammann's quantitative analysis for the reason that through the geologic ages great amounts of oxygen became "fossil," i.e., deposited in oxidized sediments. Goldschmidt was impressed by the fact that the amount of chemically combined oxygen in the lithosphere was insufficient for a complete saturation of the highest valences of the cations (silicon and metals),[2] and he concluded that the primeval oxygen soon became "fossil" and that all oxygen in the present atmosphere was formed secondarily by photosynthetic reduction of carbon dioxide.

This can be elucidated by a simple computation. The mass of the atmosphere is approximately 5.1×10^{15} tons; and the mass of free oxygen in the atmosphere is therefore only 1.15×10^{15} tons. The mass of the lithosphere (16 km. depth) is $19{,}000 \times 10^{15}$ tons; in addition to 3.1 per cent Fe_2O_3, it contains 3.71 per cent, or 705×10^{15} tons, of FeO.

Whether or not this large mass of ferrous iron is oxidizable, it is impossible to say. Part of it is present in magnetite and is not oxidizable, the largest part is combined in silicates, and the geological facts indicate that it is also not oxidizable.

However, if we make the assumption that all ferrous iron is available for oxidation according to the equation

$$2FeO + O = Fe_2O_3 ,$$

then the total free oxygen in the atmosphere would be sufficient to saturate $(1.15 \times 100)/78.5 = 1.5$ per cent of the lithosphere, corresponding to the very modest depth-range of 240 meters.

Assuming geologically stable conditions, a laterite weathering to a depth of 240 meters all over the globe is easily possible.

The attainment of this condition would deplete the air of all oxygen. Certainly, organic life is in a precarious situation. Perhaps this is now the condition of Mars with its red surface and deoxidized atmosphere. Fortunately, there are very few "fossil" or stationary situations in our good earth. Oxygen, in addition to its biochemical cycle, undergoes an equally important inorganic cycle.

Highly oxidized sediments and surface rocks are taken down to great depths by orogenic movements. Oxygen slowly re-

[2] Excerpt from Goldschmidt (1938, p. 27): "Dieses Defizit an Sauerstoff tritt zweifellos noch stärker in Erscheinung, wenn man die tieferen Teile des Erdballes mitberücksichtigt. Wir dürfen daher mit Bestimmtheit sagen, dass die Menge des Sauerstoffs im Gesamterdball nicht zur Absättigung der elektropositiven Elemente ausreicht, wobei die relativ kleine Gewichtsmenge des, sekundär entstandenen, freien Sauerstoffs der Atmosphäre vernachlässigt werden kann."

turns to the surface partly in rock minerals and partly combined in H_2O or CO_2. Likewise, atomic oxygen would seem to regenerate by the dissociation of oxides at high temperature and pressure. Then new rocks at the surface are oxidized and again reduced at depth.

In each cycle a slight separation of the oxygen isotopes takes place; O^{18} is preferentially retained in the mineral lattices of the deep-seated rocks; O^{16} is delivered into the hydrosphere and atmosphere. Isotope determinations on suitably selected material would be of great geological interest.

ACKNOWLEDGMENT.—I wish to express my thanks to Mr. Terkel Rosenqvist, of the Institute for the Study of Metals of the Univeristy of Chicago, who has contributed by discussions of several problems, and by assembling the constants given in table 3.

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OXYGEN IN ROCKS: A BASIS FOR PETROGRAPHIC CALCULATIONS

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ABSTRACT

The standard cell is defined as a rock unit containing 160 oxygens. The sum of the cations (silicon and metals) associated with this unit is very nearly 100 in all ordinary rock types. It is shown that, by using the standard cell as a basis, important petrogenetic relations can be surveyed and quantitatively defined more satisfactorily than by using any other method of petrographic calculation, including the norm method.

THE STANDARD CELL OF A ROCK

In the preceding article on the distribution of oxygen in the lithosphere it was demonstrated that the distribution follows the theoretical laws of thermodynamic equilibrium demanding a regular decrease of oxygen with depth.

This fact can be presented in two ways: in terms of decreasing oxygen percentages or in terms of increasing numbers of cations associated with a fixed number of oxygens. In table 5 of the preceding paper a rock unit containing 160 oxygens was used as a reference standard; the table demonstrates that in near-surface rocks less than 100 cations are associated with this unit and that at greater depth more than 100 cations are present. The use of the so-called "standard" cell of 160 oxygens was proposed in an earlier paper on rock alteration (Barth, 1945, 1948). In the present paper I shall offer this method for general consideration to be used as a basis in petrographic calculations.

In most rocks oxygen makes up about 92 per cent by volume; all cations taken together (silicon and metals) make up but 8 per cent by volume. Consequently, the number of oxygen ions is of the utmost importance for the volume relations in rocks. Of secondary importance is the packing, and of still less importance are the number and kinds of the constituent cations.

In petrographic calculations it is often important to compare rocks of equal volume because most replacement processes in rocks and mineral deposits take place without appreciable change in volume, often with preservation of the most delicate structures. Likewise, rock weathering, alteration by hot springs, and similar processes are known to occur without much change in volume and with the original structures and textures surprisingly well preserved.

In view of what has been said in the preceding paper about the role of the oxygen ion in rocks and in mineral lattices, it is reasonable to suppose that the mechanism of such isovolumetric alterations is that of a migration and exchange of cations in a medium composed of relatively stationary oxygen ions. Only in this way can one explain the preservation of the delicate structural features; for removal of the large oxygen ions would break up the minerals and destroy the fragile and fine structure patterns. In harmony with this idea is, e.g., the mechanism of the weathering of biotite (bleaching, baueritization), which is effected by selective removal of the metal ions while oxygen and silicon remain in the residual lattice.

Thus, if we think that rock alterations take place without change in volume, it is equivalent to saying that the number of oxygen ions has remained constant. If

we want to compare isovolumetric rock units, then we should compare units containing the same number of oxygens. These considerations in their first approximation are restricted to rocks belonging to the same mineral facies.

The leptologic structure of most silicate rocks is such that on 160 oxygens there are always very nearly 100 cations. If we list separately the cations that are associated with 160 oxygen ions, the list will, therefore, very nearly, correspond to the atomic percentages of the rock-forming elements except oxygen, or, as we shall see presently, to the so-called "equivalent molecular percentages" as introduced by Niggli (1936, pp. 295–317) some years ago.

As a standard of comparison it is expedient, therefore, to choose a rock volume containing exactly 160 oxygen ions. A volume of this size I propose to call the "standard cell of a rock."

By referring the chemical composition of a rock to its standard cell we obtain a survey of important petrological relations not revealed by a simple inspection of the analytical data.

OXYGEN AND THE ROCK-FORMING MINERALS

Before discussing the various rock types, it is important to investigate the relations of the rock-forming minerals. Pertinent data are collected in table 1. The chemical properties of the various minerals excellently illustrate the fact that oxygen and, of course, hydrogen show a strong tendency to accumulate near the surface and to become scarcer with depth.

In the minerals of the weathered sedimentary rocks few cations—about 80 or less—are associated with the standard cell, while the hydrogen content is very high; in the minerals of the deep-seated, metamorphic rocks the number of the cations in the cell attains 120, and no hydrogen is present. The cation concentration in all other mineral assemblages is between these two extremes.

This illustrates the geological cycle of oxygen in the lithosphere. We shall call it the "geological oxidation and reduction cycle." At the surface the minerals are *geologically oxidized*, which is mainly effected by the weathering agents, which remove the metal ions from the mineral lattices, often replacing them by hydrogen ions, thus entailing a relative abundance in oxygen. At great depths the minerals are *geologically reduced*, those rich in hydrogen and oxygen recrystallize, water is expelled, and a higher number of cations are introduced into the standard cell.

NIGGLI'S METHOD

In an important work on petrographic calculations, Niggli (1936) rightly emphasizes the importance of a method by which one can rapidly survey the relations between the mineralogical and the chemical composition of a rock.

The calculation of the so-called "norm" is a tool which, particularly for igneous rocks, has proved its great value. But in metamorphic rocks, in which the interrelation of mineralogy and chemistry is of the utmost importance, the norm classification has failed. The reason is obvious: The normalization of the different metamorphic mineral facies encounters great difficulties in a norm based on weight percentages. The chemical relations of a rock in terms of weight percentages obscure the comprehensive view; the computations, moreover, become unnecessarily cumbersome.

I propose, therefore, that calculation of the classical weight norm be altogether discontinued. I regard it as obsolete and

TABLE 1*

NUMBER OF CATIONS IN ROCK-FORMING MINERALS
(Referred to a Unit Cell of 160 Oxygens)

Minerals of the Weathering and Zeolite Zone		Σ Cations	H
Kaolinite†	$Al_4Si_4O_{10}(OH)_8$	71	71
Montmorillonite‡	$Y_{2-3}Z_4O_{10}(OH)_2$	80	27
Brucite	$Mg_3(OH)_6$	80	160
Zeolite	$W_{8-12}Z_{40}O_{80}\cdot sH_2O$	<80	≈120

Epimetamorphic Facies		Σ Cations	H
Chlorite	$Y_5Z_3O_5(OH)_4$	88.9	71
Talc	$Mg_3Si_4O_{10}(OH)_2$	93.3	26.7
Muscovite	$KAl_2Si_3O_{10}(OH)_2$	93.3	26.7
Zoisite	$Ca_2Al_3Si_3O_{12}(OH)_2$	98.5	12.3
Actinolite	$X_2Y_5Z_8O_{22}(OH)_2$	100	13.3
Albite	$NaAlSi_3O_8$	100	0

Dynamometamorphic Facies		Σ Cations	H
Chloritoid	$FeAl_2SiO_5(OH)_2$	91.4	45.2
Staurolite	$FeAl_4Si_2O_{10}(OH)_2$	93.3	26.7
Muscovite		93.3	26.7
Cyanite, Sillimanite	Al_2SiO_5	96.	0
Zoisite and epidote		98.5	12.3
Alk.-hornbl.	$W_3Y_3Z_8O_{22}(OH)_2$	106.7	13.3
Feldspar	WZ_4O_8	100	0
Jadeite	$NaAlSi_2O_8$	106.7	0

Mesometamorphic Facies		Σ Cations	H
Muscovite	$KY_{3-4}Z_3O_{10}(OH)_2$	93.5	26.7
Biotite		106.7	26.7
Hornblende	$X_{2-3}Y_5Z_8O_{22}(OH)_2$	100.0	13.3
		106.7	13.3
Cordierite	$Mg_2Al_4Si_5O_{18}$	97.8	0
Feldspars	WZ_4O_8	100	0
Pyroxene	XYZ_2O_6	106.7	0
Garnet	$X_3Y_2Z_4O_{12}$	120	0

Katametamorphic Facies			
Spinel	Y_3O_4	120	0
Olivine	X_2ZO_4	120	0

Magmatic Facies		Σ Cations	H
Occasionally formed in magmas			
Feldspars	WZ_4O_8	100	0
Pyroxenes	XYZ_2O_6	106.7	0
Leucite	$KAlSi_2O_6$	106.7	0
Melilite	$W_2YZ_2O_7$	114.3	0
Nepheline	$NaAlSiO_4$	120	0
Olivine	X_2SiO_4	120	0

* Quartz has the formula $Si_{16}O_{32}$ and may be present in all facies. In the table the letters W, X, Y, Z, designate ions of the following volume relations:

Ion	Si^{+4}	Al^{+3} Fe^{+3}	Mg^{+2} Fe^{+2}	Na^{+} Ca^{+2}	K^{+}
Radius	0.39	0.57 0.67	0.78 0.82	0.98 1.06	1.33
		← Z →	← Y →	← X →	← W →

† Kaolinite, dickite, halloysite, nakrite.

‡ Montmorillonite, beidellite, nontronite, hectorite, saponite.

superseded partly by the Niggli "molecular norm," partly by the norm of the standard cell, as will be introduced in the present paper.

Niggli saw the great simplification in converting weight units into molecular units which he called "equivalent units" and which are defined as follows: Equivalent formula units are those which contain the same number of the constituent cations.[1]

Obviously, the equivalent units must be defined; for if one, e.g., writes "1 Nepheline" it may mean $NaAlSiO_4$ or $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$, or $\frac{1}{3}(NaAlSiO_4)$, etc.

In $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$ the sum of the cations would be $2Na + 2Al + 2Si = 6$. The "equivalent" unit of forsterite would be $4MgO \cdot 2SiO_2$ ($4Mg + 2Si = 6$); and of enstatite, $3MgO \cdot 3SiO_2$ ($3Mg + 3Si = 6$). It is simplest to reduce the several equivalent units to a sum of the cations = 1, then the sum of the several cations represents the number of equivalent units, thus (using the customary abbreviations, like Ne for Nepheline, etc.): $Na_2O \cdot Al_2O_3 \cdot 2SiO_2 = 6Ne$; $3MgO \cdot 3SiO_2 = 6En$; $MgSiO_3 = 2En$, etc.[2]

In this way the mineral equations become very simple, for instance,

$$Mg_2SiO_4 + SiO_2 = 2MgSiO_3 ,$$

$$3Fo + 1Q = 4En .$$

It is important to note that the sum of the coefficients of reaction on both sides of the equation is the same, for instance,

$$\underbrace{3Ne + 2Q}_{5} = \underbrace{5Ab}_{5}$$

Consequently, the sum of the molecular amounts remains constant. If a percentage distribution has been assumed, then any recalculation in terms of other combinations can be made without affecting the sum, which remains 100. This pertains to the norm calculation also. If a rock analysis is recalculated to a sum of 100 cations, then the molecular norm can be computed in practically no time. The molecular norm differs but slightly from the weight norm. This arises from the fact that the constituent oxides, SiO_2, $\frac{1}{2}Al_2O_3$, CaO, etc., have approximately the same weight. (Every petrographer knows, for instance, that the composition of a feldspar expressed in terms of Or, Ab, and An comes out with about the same figures, whether molecular percentage or weight percentage is used.)

Again I repeat my plea for using molecular units. An increasing number of petrographers in America and Europe alike, when publishing an analysis, calculate the corresponding norm values. It would be of great value if they would always apply molecular percentages. The fact that the molecular values are rather similar to the weight values makes the change easier.

HOLMQUIST'S METHOD

A method that should not go into oblivion but can be used with advantage in modern studies is the molecular classification introduced by Holmquist in his great study of Swedish granites (1904–1905, pp. 78–269). At a time when the normative classification was introduced by the four leading American petrologists (Cross, Iddings, Pirsson, and Washington), Holmquist had made his own system. His principal point was that the chemical relations of a magma cannot be adequately discussed on the basis of weight percentages. In the treatment of chemical compounds and chemical reac-

[1] "Als übereinstimmende Formelgrössen können wir diejenigen bezeichnen, welche die gleiche Anzahl der wichtigen elektropositiven Elemente der Gruppe Si, Al, Fe, Mn, Mg, Ca, Na, K, Ti, Zr, Cr usw. enthalten" (Niggli, 1936, p. 296).

[2] Observe that in this scheme the symbol Ne has a quantitative meaning. It does not just mean the composition $NaAlSiO_4$ or $Na_2O \cdot Al_2O_3 \cdot 2SiO_2$, but it means a quantity of this composition corresponding to one cation.

tions, as well as in petrographic calculations, the weight units must be converted into molecular units.

More than thirty years before Niggli's publication, Holmquist recommended presentation of the rock analyses in terms of the constituent cations (metals and silicons) recalculated to 100 per cent. Like Niggli, he left oxygen out of consideration. The composition of a granite, for instance (from Halen, Sweden) may be written thus:

TABLE 2

Holmquist's Comp.	Mol. Per Cent	Modal Comp.	Weight Per Cent
Quartz	32.5	Quartz	34.2
Feldspar	63.0	Feldspar	58.7
Enstatite	1.2	Biotite	3.7
Fe, Al	3.4	Magnetite	1.2

$$Si_{69.5}\,Al_{16.0}\,Fe_{1.4}\,Mg_{0.6}\,Ca_{1.4}\,Na_{5.2}\,K_{6.0}\,.$$

As the next step the cations can be rearranged as follows:

$$Si_{32.5} = 32.5\,,$$

$$Si_{36.4}\,Al_{14.0}\,Ca_{1.4}\,Na_{5.2}\,K_{6.0}\,(+Al_{2.0}) = 63.0\,,$$

$$Si_{0.6}\,Mg_{0.6}\,Fe_{1.4} = 1.2\,.$$

This grouping is self-explanatory and give an idealized normative composition which compares well with the actual, or modal, composition, as seen in table 2.

THE COMPOSITION OF THE STANDARD CELL

From the chemical analysis of a rock it is easy to calculate the elemental composition of the standard cell—it is simply to list the cations associated with 160 oxygen ions. The sum of the cations is near 100, and the figures, therefore, are nearly identical to the equivalent molecular percentages of Niggli and, furthermore, rather similar to the weight percentages as given by the chemists. This is important, because petrologists are accustomed to judge the composition of a rock by simply inspecting the weight percentages of the oxides. The numbers expressing the elemental composition of the standard cell are sufficiently close to those expressing the weight percentages of the oxides to allow us, generally speaking, to go by the old standards. This can be seen by inspection of table 3, which gives an example of the chemical composition of a hornblende minette expressed as (1) weight percentages of the oxides, (2) equivalent molecular percentages, and (3) number of cations associated with the standard cell. Table 4 gives, furthermore, the normative composition of the same rock in terms of (1) the usual

TABLE 3*

	Weight Per Cent	Eq. Mol. Per Cent	No. of Cations in Standard Cell
SiO_2	52.68	51.6	49.4
TiO_2	2.71	2.0	1.9
Al_2O_3	14.02	16.2	15.6
Fe_2O_3	5.04	3.7	3.5
FeO†	6.40	5.3	5.1
MgO	3.35	4.8	4.6
CaO	6.42	6.8	6.5
Na_2O	2.43	4.6	4.4
K_2O	3.30	4.1	3.9
P_2O_5	1.30	1.0	1.0
H_2O	1.89	(6.2)	(5.9)
S	0.60	(1.0)	(1.0)
Sum	100.14	100.0	95.9
− for S_2	0.15		

* The procedure of calculation is as follows: The figures giving the weight percentages are divided by the equivalent molecular weight of the corresponding oxide. Thus the figure for SiO_2 is divided by 60.06, the figure for MgO is divided by 40.32, and so on for all oxides containing one cation in the formula. For oxides containing *two* cations in the formula, we must divide by *one-half* the formula weight. Thus the figure for Al_2O_3 is divided by 50.97 (= ½101.97), the figure for K_2O is divided by 47.1 (= ½94.2), etc. Thus we arrive at the cation proportions which, recalculated to 100 per cent, give the equivalent molecular percentages. Likewise, the figures for the proportions of the cations may be recalculated to give directly the number of cations in the standard cell. One just has to remember that there are ½, 1, 1½, or 2 oxygen ions associated with each individual cation and that the sum of the associated oxygens should be 160.

† Including 0.15 per cent MnO.

weight norm, (2) the Niggli molecular norm, and (3) mineral molecules in the standard cell. It is evident that the three different norms equally well accord to the petrologist a rapid survey of the potential mineral composition of the rock.

The "norm" of the standard cell as shown in the fourth column of table 4 is

TABLE 4

	Weight Norm	Eq. Norm	Norm of Standard Cell
Q	11.1	10.5	10.1
Or	19.6	20.5	19.6
Ab	21.1	23.0	22.1
An	18.2	18.7	17.9
Wo	2.8	2.8	2.7
Fs	3.1	2.7	2.6
En	8.4	9.6	9.2
Ap	2.8	2.6	2.5
Il	5.3	4.0	3.8
Mt	7.6	5.6	5.4
Total	100.0	100.0	95.9

directly derived from the figures representing the cations in the cell (fourth col., table 3). Not only the norm but any mineral combination that one might like to express, normative or modal, can be exhibited by simply recombining the same figures. This advantage the standard-cell method shares with the Niggli method. In this respect the classical-norm method utterly fails.

How the cations of the standard cell can be recombined to give, quantitatively, the actual, rather complicated mineral composition of the rock, as qualitatively determined with the microscope, is demonstrated in table 5.

The chemical "formula" of the rock (analogously to the Holmquist method) can be presented as follows:

However, the standard-cell formulas are, in one respect, fundamentally different from those proposed by Holmquist; they are based on a unit containing 160 oxygen atoms, whereas Holmquist disregarded the oxygen. Niggli likewise disregarded the oxygen, and therefore neither of these two methods can reflect the volume relations, which are of fundamental importance in geology.

EXAMPLE 1: DIABASE—HORNBLENDE MINETTE

The geological relations of the hornblende minette, the composition of which has been rendered in tables 3 and 4, present petrogenetic problems of general interest. The minette occurs 20 km. west of Kristiansand, on the southern coast of Norway (Barth, 1942). The mode of occurrence of the minette is that of a diabase dike apparently cutting pre-Cambrian migmatite of granitic composition; but a closer inspection shows that part of the dike is again cut by a pegmatitic facies of the migmatite. Consequently, the dike represents an old lamprophyre of pre-Cambrian age manifestly older than the last phases of the pre-Cambrian orogeny.

The mineral composition of the minette is obviously secondary, i.e., not magmatic but acquired through recrystallization. The analysis indicates that chemical alterations may have taken place at the same time. Briefly stated, the genetic problem is to investigate and define the metasomatic changes that took place during the metamorphism.

Without adducing further evidence it can be stated that there are reasons for

$$K_{3.9}\,Na_{4.4}\,Ca_{6.5}\,Mg_{4.6}\,Fe_{8.6}\,Al_{15.6}\,Ti_{1.9}\,Si_{49.4}\,P_{1.0}\,[O_{148.2}\,(OH)_{11.8}]_{160}\,.$$

believing that the rock represents a diabase of the usual (basaltic) type, which later was changed into a hornblende minette.

Diabase dikes exhibit a remarkably homogeneous composition. Without any great error, therefore, we can assume that the pristine composition of the hornblende minette corresponds to the average analysis of all diabase dikes of the pre-Cambrian of southern Norway. rographic methods are used. Some petrographers introduce entirely arbitrary assumptions, such as that iron or alumina is constant during the metasomatism. Now most petrologists agree that metasomatic changes take place volume for volume—and in some cases laborious calculations have been carried out to give educt and product the same volume. In the new method for calculation here proposed, the volume relations are auto-

TABLE 5

Si	Ti	Al	Fe'''	Fe''	Mg	Ca	Na	K	P	(S)	(H_2O)	Modal Minerals
49.4	1.9	15.6	3.5	5.1	4.6	6.5	4.4	3.9	1.0	(1.0)	(5.9)	95.9
13.8	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	13.8 Q
7.5	.....	2.5	.....	.....	.....	.....	.....	2.5	.....	.....	.....	12.5 Or
12.1	.....	4.1	.....	.....	.....	.....	4.1	.....	.....	.....	.....	20.3 Ab
2.5	.....	2.4	.....	.....	.....	1.2	.....	.....	.....	.....	.....	6.1 An
7.8	0.5	3.2	0.7	2.2	2.2	2.4	0.3	.....	.....	.....	(1.2)	19.3 Ho
3.5	0.2	2.2	0.6	1.9	1.0	.....	.....	1.4	.....	.....	(1.4)	10.8 Bi
1.0	.....	1.2	.....	0.2	1 4	.....	.....	.....	.....	.....	(1.5)	3.8 Chl
1.2	1.2	.....	.....	.....	.....	1.3	.....	.....	.....	.....	.....	3.7 Ti
.....	.....	.....	.....	.....	.....	1.6	.....	.....	1.0	.....	.....	2.6 Ap
.....	.....	.....	2.2	0.8	.....	.....	.....	.....	.....	(1.0)	(1.7)	3.0 Ore

Consequently, a comparison of the chemical composition of the original diabase with that of the minette should give us a clue to the changes. Such a comparison is difficult, however, if conservative pet- matically satisfied by referring everything to the standard cell. We compare the two standard cells directly, and the problem is solved without any trouble at all.

DIABASE[3]

$$K_{1.4}\,Na_{4.5}\,Ca_{9.0}\,Mg_{5.9}\,Fe_{10.7}\,Al_{16.0}\,Ti_{1.9}\,Si_{48.5}\,P_{0.2}\,[O_{150.6}\,(OH)_{9.4}]_{160}\,.$$

MINETTE

$$K_{3.9}\,Na_{4.4}\,Ca_{6.5}\,Mg_{4.6}\,Fe_{8.6}\,Al_{15.6}\,Ti_{1.9}\,Si_{49.4}\,P_{1.0}\,[O_{148.2}\,(OH)_{11.8}]_{160}\,.$$

Since the two formulas represent the contents of the standard cells which have approximately the same volume, the difference between the rocks is found simply by subtracting the one from the other. Thus it is seen that the original diabase is metasomatically transformed into hornblende minette by addition and subtraction of the cations given in table 6.

Table 6 demonstrates that a remarkably small fraction of the rock substance (less than 1 per cent) need migrate in order to effect great metasomatic changes.

[3] Diabase dikes cutting pre-Cambrian rocks, southern Norway (average of 8). Recalculated from Barth-Correns-Eskola (1939, p. 72, table 24).

EXAMPLE 2: SOLFATARIC ALTERATION

In a chapter on "Petrography of Hawaii," Macdonald (Stearns and Macdonald, 1946, p. 192) describes solfataric alteration at Kilauea Caldera thus:

> Gases escaping along faults at Kilauea Caldera have altered the adjoining rocks, both lavas and tuffs. Most prominent is a type of alteration caused by steam containing low concentrations of sulfuric, sulfurous, and carbonic acid. The resultant product is a rock composed largely of opal, with smaller amounts of kaolinite or related clay minerals, and relict magnetite and ilmenite. It has approximately the same volume as the unaltered rock, and original structures and textures are surprisingly well preserved.

These observations indicate that the number of oxygens must have remained constant during the alteration. Again we compare the standard cell of the parent-rock with that of the decomposed rock (recalculated from Payne and Mau, 1946, p. 351).

TABLE 6

DIABASE PASSES INTO HORNBLENDE MINETTE BY

Adding	Subtracting
2.5 ions of K	0.1 ions of Na
0.9 ions of Si	2.5 ions of Ca
0.8 ions of P	1.3 ions of Mg
2.5 ions of H	2.1 ions of Fe
	0.4 ions of Al
Total: 6.7 cations representing 12.6 valences	6.4 cations representing 12.6 valences

PARENT

$$K_{0.5}\ Na_{3.7}\ Ca_{10.6}\ Mg_{8.4}\ Fe_{7.0}\ Al_{17.0}\ P_{0.2}\ Ti_{1.8}\ Si_{48.8}\ O_{160.0}\ H_{8.4}\ .$$

DECOMPOSED

$$--\ Ca_{0.1}\ -\ Fe_{0.2}\ Al_{0.1}\ -\ Ti_{8.4}\ Si_{55.9}\ O_{160.0}\ H_{69.6}\ .$$

Figure 1 illustrates graphically the changes indicated by the chemical analyses. Figure 1, *A*, is copied after Macdonald and Payne and Mau (1946, p. 351, table 3), plotted on the basis of no change in the silica content. This assumption is, of course, quite arbitrary; and figure 1, *B*, is believed to present a truer picture of what actually happened, plotted on the basis of no change in the oxygen content. It gives the following picture of the mechanism of the alteration: The effect of the acid hot-spring water is a selective removal of all metal ions and a simultaneous introduction of hydrogen, silicon, and titanium ions. Petrographic investigations in other hot-spring areas support this mode of interpretation.

EXAMPLE 3: THE STAVANGER AREA

The injection metamorphism of the Stavanger area, southern Norway, has become well known through Goldschmidt's work in 1920 and by the fact that several international excursions to the area have been organized. The main point of petrologic interest is the progressive metamorphism of the sediments by granitic and trondhjemitic intrusions.

The sediment by low-grade metamorphism is changed into a quartz-muscovite-chlorite phyllite; by increasing metamorphism the sediment is changed into albite porphyroblast schist and, further, into a granitelike rock. Goldschmidt gives the following discussions of the transition sediment⟷ albite porphyroblast schist (pp. 113–114, translated from German):

> In order to investigate the various possibilities of material transport we can calculate the composition of various rock mixtures, the soda content of which would correspond to that of the albite porphyroblast schist.
>
> In this calculation the initial composition

is taken as the quartz-muscovite-chlorite phyllite.

The possibilities are as follows:

a) The material was introduced as trondhjemite; to 100 parts phyllite, 150 parts trondhjemite must be added.

b) The material was introduced as granite aplite; to 100 parts phyllite, 400 parts granite aplite must be added.

c) The material was introduced as albite; to 100 parts phyllite, 80 parts albite must be added.

d) The soda was introduced as albite; simultaneously silica and lime were introduced; water was removed. On 100 parts phyllite, 30 parts albite, 19.4 parts SiO_2, 3.9 parts Ca were added, 1.3 parts of water were subtracted.

e) The material was introduced and was *fixed* in form of oxides of SiO_2, Na_2O, CaO; water was removed; to 100 parts phyllite, 26 parts SiO_2, 3.1 parts CaO, 2.8 parts Na_2O were added, 1.7 parts water were subtracted.

The results of these calculations prove that the introduction of soda did not occur as a simple addition of trondhjemite, granite, or albite; but that, according to the possibility *e*, from solutions or vapors lime and soda were selectively absorbed, with simultaneous precipitation of silica.

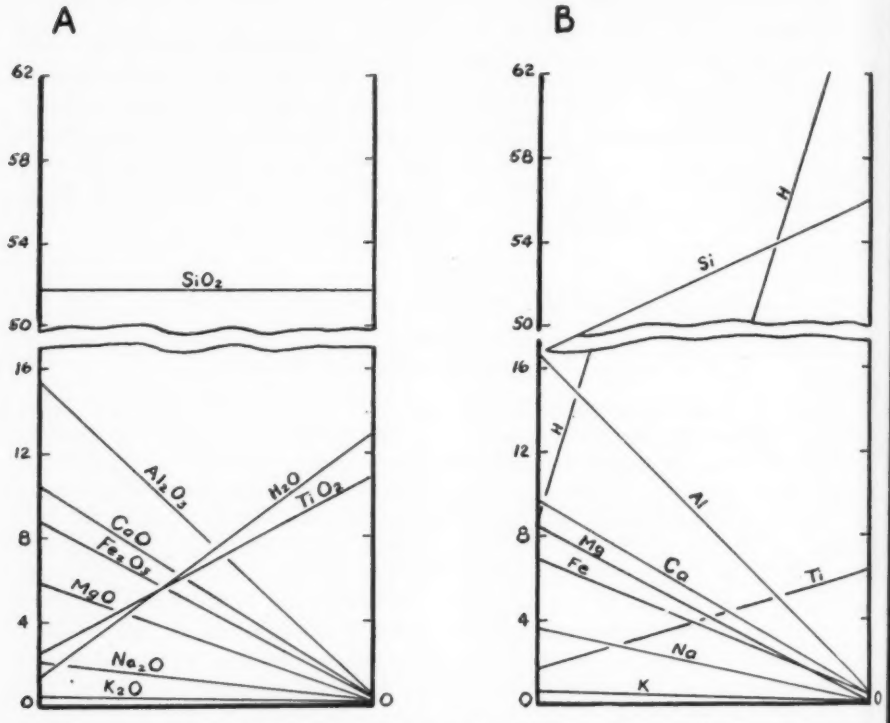

FIG. 1.—*A*, Alteration diagram according to earlier investigators; SiO_2 assumed to remain constant. *B*, Alteration diagram. Volumes are kept constant by keeping oxygen constant.

The foregoing paragraphs sum up the main conclusions and the main results of the paper by Goldschmidt, which at that time became well known and much discussed because it emphasized the importance of the functions exercised by the metasomatic solutions: the chemical composition of a metamorphic rock does

not generally correspond to a mixture of the sediment plus the injected "magma"; but from the magma certain oxides were selectively absorbed and fixed in the sediment, other oxides passed by without being fixed.

The volume relations were the chief reasons for Goldschmidt's rejecting the explanation of a simple mechanical mixing of the sediment and the intruded material. It is seen that the possibilities listed under *a* or *b* imply great quantities of foreign material to be introduced into the sediment. Goldschmidt's solution requires smaller quantities, although they are still appreciable. His solution can be represented by the following equation:

$$100 \text{ sediment} + 26.0\ SiO_2 + 3.1\ CaO + 2.8\ Na_2O - 1.7\ H_2O = \text{Albite-porphyroblast schist}.$$

Since the metamorphosed sediment shows no evidence of expansion, no swelling of the beds, etc., Goldschmidt's solution, which implies an increase of volume of more than 25 per cent, is not satisfactory. A metasomatism volume for volume is more in accordance with the field evidences. Therefore, we should compare the composition of the standard cells:

Sediment: $K_{5.1}\ Na_{2.1}\ Ca_{0.6}\ Mg_{2.4}\ Fe_{5.2}\ Al_{20.8}\ Ti_{0.6}\ Si_{51.0}\ C_{0.5}\ O_{160.0}\ H_{23.8}$.

Schist: $K_{3.9}\ Na_{5.3}\ Ca_{2.8}\ Mg_{1.9}\ Fe_{4.0}\ Al_{16.1}\ Ti_{0.4}\ Si_{57.1}\ C_{1.0}\ O_{160.0}\ H_{9.8}$.

Thus sediment passes into schist as shown in table 7.

TABLE 7

Adding	Subtracting
3.2 Na-ions	1.2 K-ions
2.2 Ca-ions	0.5 Mg-ions
6.1 Si-ions	1.2 Fe-ions
0.5 C-ions	4.7 Al-ions
	0.2 Ti-ions
Sum: 12.0 metal ions (34 valences)	7.8 metal ions and 14.0 H-ions (34 valences)

Table 7 shows that it is not sufficient to add soda, lime, and silica to the sediment to convert it to schist, as Goldschmidt claimed; there has to be a corresponding subtraction of other constituents.

Again the calculation of the chemical composition of the standard cells of the rocks gives us a much better understanding of the geological processes than does any other method of petrographic calculation.

CONCLUSIONS

The standard cell of a rock is defined as a rock unit containing 160 oxygens. In most rocks oxygen makes up about 92 per cent by volume; the cations (silicon and metals) make up but 8 per cent by volume. Consequently, the number of oxygen ions is of great importance for the volume relations in rocks. If we want to compare isovolumetric rock units, then we should compare units containing the same number of oxygens.

In an average rock very nearly 100 cations are associated with the standard cell; consequently, a list of the cations will very nearly correspond to the atomic percentages of the rock-forming elements except oxygen.

The cations thus listed may be combined to form mineral molecules belonging to any one of the mineral facies (including, of course, the mineral molecules belonging to the conventional norm); any mineral combination that one might like to express, normative or modal, can

be exhibited by simply recombining the same figures.

It is a great simplification to use molecular units rather than weight units; therefore, the weight percentages should never be taken as basis for petrographic calculations. Equivalent molecular percentages are better to use; but, generally speaking, the best basis in petrographic calculations is the list of cations associated with the standard cell; this list represents the "chemical formula" of the rock, and on the basis of this list important petrologic relations can be worked out: (1) Degree of geologic oxidation. The cations in the standard cell, which—with the designation here adopted, correspond in number to the mineral molecules formed by combination of the cations—add up to a figure close to 100. In rocks of near-surface relations the number will usually be less than 100 (= more oxygen per cation: highly oxidized rocks), in deep-seated rocks the number will be more than 100 (less oxygen per cation: low degree of oxidation). (2) Most metamorphic and metasomatic processes in rocks take place without change in volume. In order to study such processes, for instance, in the study of compositional changes induced in a rock series through progressive metamorphism, the volume relations are automatically satisfied by referring the composition to the standard cell. By directly comparing the standard cells of the rocks under investigation, one obtains a rapid review of addition and subtraction of material caused by the petrogenetic processes. In this way the mechanism of granitization, the formation of "basic fronts," and similar phenomena can be studied more effectively than by any other method of petrographic calculation.

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ISOTOPE RATIOS: A CLUE TO THE AGE OF CERTAIN MARINE SEDIMENTS

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ABSTRACT

If an element A has a radiogenic isotope, A_1, and a nonradiogenic one, A_2, the ratio A_1/A_2 is an index of the age of marine chemical sediments, if the content of the isotope B^* producing A_1 can be neglected. It is shown that the method can be used for strontium (and perhaps for Pb^{208}) in limestones and anhydrites.

INTRODUCTION

Age determinations have hitherto been performed on igneous rocks only; and this is unfortunate, as the relative geological time-scale is based on the sedimentary rocks, with their content of invertebrate fossils. Hence it may be of interest to attempt to develop methods suitable for sedimentary rocks.

PRINCIPLES OF THE NEW METHOD

If a radioactive isotope generates an isotope of another element, the abundance of the isotopes of the daughter-element will vary with time, the abundance of the isotopes, and the disintegration constant of the radioactive isotope. If we consider the earth's crust, the isotopic composition is a special, evaluable function of time; but this statement is not true for a small part of it. The function will depend on the composition of the sample, its geological history, etc.

Let us now look at the problem from a different point of view. We can ask if there is any geological milieu about which we can be certain that the isotopic composition of an element is a one-valued function of time. I think that the sea is such a milieu. The isotopic composition of sea water is a one-valued function of time and probably represents, to a high degree of approximation, an average sample of the earth's crust.

If there is a difference between the earth's crust and the sea, the maximum would exist under the very special topographical conditions during which large fractions of the land areas do not drain to the sea. But even in this case I think the approximation to the earth's crust must be very close.

If a chemical sediment is precipitated in the sea, the isotopic constitution of the radiogenic element will also be a one-valued function of its age, if the content of the radioactive parent can be neglected. This means that the isotope ratios, at least in principle, can be used as age indicators and that the index multiplied by a constant gives the age of the sediment in millions of years.

The most suitable sediments seem to be limestones, anhydrites, and gypsum. In these cases strontium and lead are the most promising elements. The other disintegration products, such as helium, argon, calcium, neodymium, hafnium, and rhenium, are of no significance in this connection. The special problems of strontium and lead will be discussed separately.

STRONTIUM IN LIMESTONES AND ANDHYDRITES

Strontium should be particularly favorable, as it is an important constituent of sea water, 13,000 γ/liter; and, on the other hand, the content of rubidium, which is the parent, is low, namely, 200

γ/liter (Sverdrup, Johnson, and Fleming, 1942). As W. Noll has shown (1934), strontium can replace calcium in calcium minerals such as calcite, dolomite, aragonite, gypsum, and anhydrite; and rubidium is, on the other hand, known not to enter these minerals. Strontium occurs most frequently in aragonite and anhydrite (table 1).

TABLE 1

VARIATION IN CONTENT OF STRONTIUM IN CERTAIN SEDIMENTARY MINERALS AND ROCKS ACCORDING TO W. NOLL (1934)

Mineral, Rock	Per Cent SrO
Gypsum	0.003 –(0.13)
Anhydrite	0.17 – 0.69
Calcite	0.0005– 0.14
Aragonite	Up to 4.69

TABLE 2

STRONTIUM CONTENT OF RECENT MARINE ORGANISMS, AFTER NOLL (1934)

	Per Cent SrO
Corals:	
Porites clavaria, Florida	0.5
Corallium rubrum, Sicily	0.2
Millepora alcicornis, Florida	0.5
Cephalopod:	
Nautilus pompilius, Pacific Ocean	0.4
Brachiopod:	
Terebratula vitrea, Naples, Italy	0.05
Lamellibranchiata:	
Pinna squamosa, Naples, Italy	0.04
Gastropod:	
Bulla ampulla, Atlantic Ocean	0.19
Scaphopod:	
Dentalium, North Sea	0.2
Lithothamnium:	
Lithothamnium polymorphum, Adriatic Sea	0.3

The high strontium content of gypsum refers to rock-forming gypsum, which is a product of metamorphism of anhydrite. Noll found that the content of strontium in pure gypsum crystals was very low. Thus it is probable that strontium occurs as celestine in rock-forming gypsum. In recent marine organisms the content of strontium can rise relatively high, as table 2 shows.

From these figures it must be concluded that the shells themselves can be used for age determinations. In this case the question arises: Are the strontium isotopes selectively concentrated by organisms? I think it is very improbable that a biological separation in this case would be of any importance at all; but in any case it is simple to investigate shells of recent marine organisms. Celestine is regularly found associated with fossils which have had shells of aragonite. Noll cites several examples from the literature (1934, pp. 574–575). Thus we must discuss one of the serious problems confronting this method. Are the isotope ratios changing during the diagenesis by the mixing of strontium atoms of different origin? This question can be answered only by an investigation of a thick lime deposit; but, nevertheless, it is of interest to give the opinion of Noll, especially in the case of crystals of celestine associated with fossils. He says (1934, pp. 575–576):

Zu demselben Resultat: Verknüpfung der Coelestinanreicherungen mit Aragonittieren kam Liebetrau bei Untersuchungen in Thüringer Muschelkalk. Coelestin vertritt nach ihm ganz besonders gern die Substanz von Gastropodenschalen, nur untergeordnet von Lamellibranchiaten und nicht von Brachiopodenschalen. Es sind aber eben Gastropoden und Lamellibranchiatenschalen, die aus Aragonit bestehen. Liebetrau sieht die Erklärung für die Verknüpfung des Coelestins mit den Aragonithartteilen in der grösseren Löslichkeits des Aragonites gegenüber Calcit und damit der grösseren Umsatzfähigkeit des Aragonites mit zirkulierenden Lösungen. Damit dürfte aber noch nicht alles erklärt sein. Durch die Auflösung des Aragonites kann zwar ein Hohlraum entstehen, in dem Coelestinkristalle sich frei ausbilden können, *dass* sie

sich aber abscheiden, muss einen anderen Grund haben, der nach dem Massenwirkungsgesetz entweder in Erhöhung der Sr^{2+} oder des SO_4^{-2}-Ionenkonzentration gesucht werden muss. Die Konzentration des Sr^{2+} kann offenbar im Bereich der Auflösung der Aragonitschale erhöht werden, da die Aragonitteile Strontium angereichert enthalten. Der Angelpunkt für die Erklärung der Verknüpfung von Coelestin mit Aragonitschalen wäre also darin zu sehen, dass der hohe Strontiumgehalt des Aragonites durch dessen leichte Löslichkeit leicht freigemacht und im entstandenen Lösungshohlraum sogleich wieder abgeschieden werden kann.

Hence Noll concludes that there is very little mixing of material. This is, of course, especially the case when thin beds of limestone are separated by shales or other dense materials. Whether the method can be used will depend on the magnitude of the variations of the isotope ratios. In this respect we must know the content of Rb^{87}, Sr^{87}, and Sr^{86} (or another nonradiogenic isotope of strontium) in the earth's crust, and the disintegration constant of Rb^{87}.

According to V. M. Goldschmidt (1937*b*), the content of rubidium is 310 gm/ton, and 28 per cent of this amount is radioactive rubidium 87. Thus the content of Rb^{87} in the earth's crust is 86 gm/ton. The disintegration constant is, according to a new determination by S. Eklund (1946), $3.8 \pm 0.7 \times 10^{-19}$ sec^{-1}.

The strontium content in the earth's crust is much more difficult to estimate. According to Noll (1934), the content of SrO would be 0.05 per cent. His estimate was not made according to the common method of estimating the abundances of elements. Goldschmidt (1937*a*) uses the value 150 gm. Sr per ton. This value is probably low; a plausible value is 200–300 gm. Sr per ton. Using the higher value for our calculation, 300 gm. Sr per ton, and an isotope abundance of Sr^{87} of 6.6 per cent, the present content of Sr^{87} in the earth's crust would be 20 gm/ton. The present ratio Sr^{87}/Sr^{86} is arbitrarily fixed as 0.6600. In figure 1 the strontium ratio is plotted on a geological time-scale. The lengths of respective periods are given according to A. Holmes (1947).

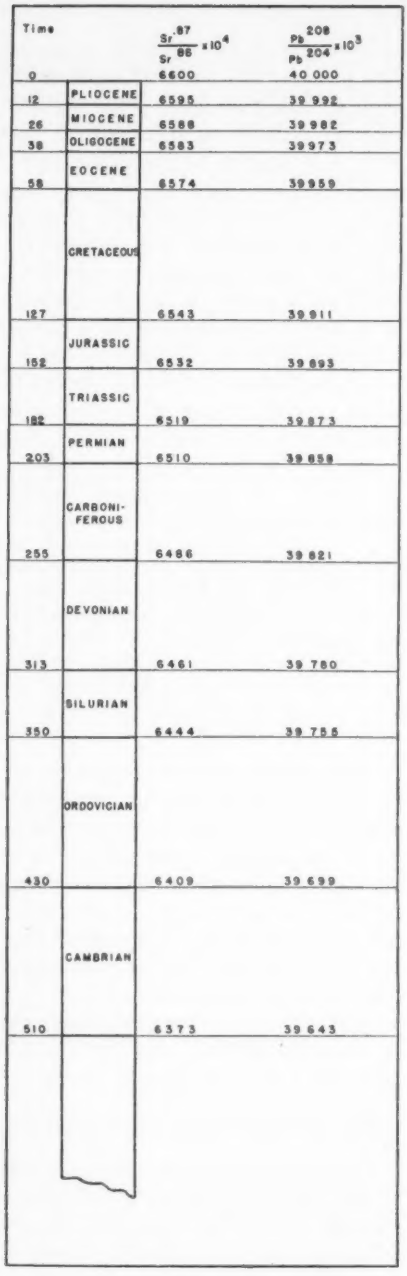
From this figure it is quite clear that the measurement of the isotope ratio Sr^{87}/Sr^{86} must be made as accurate as possible. In order to get the opinion of an expert, I have discussed the problem with Dr. Alfred C. O. Nier, of the University of Minnesota, and he writes, in a letter to the present writer (published with the kind permission of Dr. Nier):

> There is little doubt in my mind that variations of 0.1% in the ratio can be detected if sufficient care is taken, and in time the precision may be extended to permit measurements to within 0.01%. Certainly there is no fundamental reason why a mass spectrometer could not be perfected to give such accuracies. Naturally, for some time to come it will be difficult, if not impossible, to achieve an accuracy of 0.01% on a routine basis, but if enough effort is expended it should be possible to obtain this accuracy as a regular thing.

This means that in the future the ratio can be given to four places, and the error will be about 0.00006. In the present case this is equivalent to the statement that the ages can be given with an error of about 1,000,000 years, regardless of their position on the time-scale. This error refers to the relative error and the relative time-scale, and probably it will be very difficult to determine the relation between the absolute age and the strontium ratio with the same accuracy.

LEAD IN LIMESTONES AND ANHYDRITES

If the strontium content of sea water is high, the lead content is very low. According to M. Boury (1938), the lead content is about 4 γ/kg sea water. The uranium content, on the other hand, is 1.5 γ/kg, and the thorium content much



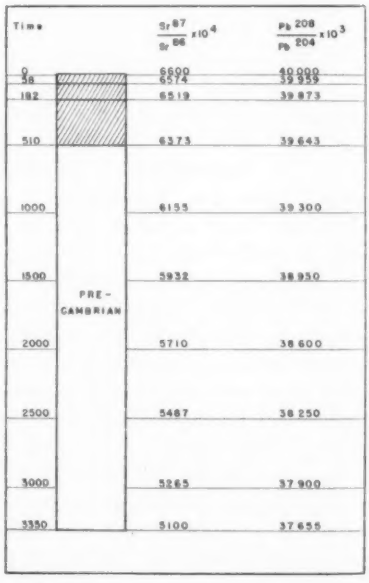


FIG. 1.—The variation of the ratios Sr^{87}/Sr^{86} and Pb^{208}/Pb^{204} with time

less than 0.5 γ/kg. These two measurements were made by H. Pettersson and his co-workers (1939).

Measurements of the lead content of marine sediments are very rare. According to Siebenthal (1915, pp. 78–79), in Cambrian and Ordovician magnesian limestones in Missouri the lead content varies between less than 4 to 16 gm/ton, with a mean of 10 gm/ton. In Mississippian limestones in Missouri, on the other hand, the content was about 20 gm/ton and in different limestones and dolomites from Iowa about 32 gm/ton. Forchhammer[1] found in 1865 the lead content in corals to be as follows:

	Gm/Ton
Pocillopora alcicornis	2.7
Heteropora abrotanoides	20

Of course, these figures are not of any great value, but they fit into what can be expected. As the ionic radii of Sr^{2+} and Pb^{2+} are very similar, it is probable that the lead and strontium content in sea water and marine chemical sediments are proportional. If this is the case, the lead content of limestones must be expected to be up to 10 gm/ton and in anhydrites up to 1 gm/ton.

Hence this method can be dangerous if the content of radioactive elements is relatively high. For electrostatical reasons it is not probable that uranium and thorium can occur in the crystal structure of calcite, aragonite, anhydrite, and gypsum. This is also in agreement with experience; Beer and Goodman write, for instance (1944, p. 1251):

> A pure, crystalline limestone or a pure, white quartz sand exhibits practically no measurable radioactivity. Those limestones which show appreciable activity also contain proportionate amounts of shale, organic matter, or other material which has once occupied the colloidal phase.

[1] Cited in Boury (1938).

Without any opinion on the correctness of their theory, it is nevertheless clear that the radioactivity of pure marine limestones is low. Probably it is advisable to use only the ratio Pb^{208}/Pb^{204}, that is, the ratio referring to the thorium series, as the thorium content of sea water is extremely low and the presence of organic matter can be an indication of an enrichment of uranium.

We assume the thorium content of the earth's crust to be 4 gm/ton and the dis-

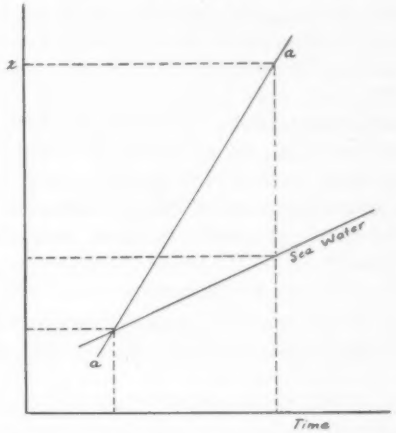


FIG. 2.—The variation of the ratio $Pb^{206}/Pb^{204} = x:1$ with time in limestone and sea water.

integration constant to be 4.99×10^{-11} $year^{-1}$. The lead content is 20 gm/ton, and the percentage of Pb^{208} is about 52. Thus the amount Pb^{208} can be estimated to be 11.5 gm/ton. These values give the ratios Pb^{208}/Pb^{204} plotted in figure 1.

Assuming the same accuracy as in the case of strontium, the error can be estimated to be about 7,000,000 years.

LIMESTONES WITH HIGH CONTENT OF URANIUM

If the limestones contain uranium, then the present method cannot be used. On the other hand, the usual lead method

cannot be used, as the original composition of the common lead is not known. But, with some modifications, it is possible to use a method proposed by the present writer several years ago (1943). The principle is shown in figure 2. If we determine the isotope ratio of the limestone, we get the ratio $1:x$. From the content of uranium and lead in the limestone, we can recalculate the variation of the isotope ratio with time (curve *a*). On the other hand, we know that the isotopic composition of the sample at the time of deposition was the same as the isotope composition of the earth's crust. This last curve is also known, and the point of intersection gives the age. If this method is to be accurate, the ratio of uranium to lead must be either high or low in relation to the ratios in the earth's crust, as can easily be seen from the figure.

CONCLUSIONS

1. It is shown that the ratio between a radiogenic and a nonradiogenic isotope of an element can be used as an index of the age of marine chemical sediments, if the content of the mother-element can be neglected.

2. The method is especially suitable for strontium in limestones and anhydrites. It will be possible to determine the ratio with an error of 0.01 per cent. In order to get accurate absolute age determinations, the contents of rubidium and strontium in the earth's crust and the disintegration constant of Rb^{87} must be determined with greater accuracy. The error of an age determination can be estimated to be 1,000,000 years.

3. It might be possible to use lead in some cases to check the results obtained with strontium ratios. The accuracy to be expected is less than in the case of strontium, perhaps 7,000,000 years.

4. A variation of the method proposed by the present writer (1943) is shown to be of some value in the determination of the age of marine limestones either rich or poor in uranium.

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THE PRESERVATION OF ANTARCTIC ICE SPECIMENS[1]

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ABSTRACT

Small ice specimens, including thin sections and delicate ice crystals, were brought back intact from the Antarctic. The samples were kept in airtight containers and were stored in a cold-storage room aboard ship during the journey home. Thin sections were housed in slide boxes in which the sections rested horizontally. Immersion of the sections in kerosene to prevent evaporation was not entirely satisfactory because of the tendency of the kerosene to penetrate the section. However, such penetration can be prevented. The use of cover glasses with a seal of vaseline around the edges was found by European glaciologists to prevent evaporation. Specimens other than thin sections were packed in snow. The snow prevented jostling and served as an insulating medium.

The feasibility of transporting small specimens of ice over distances of thousands of miles in journeys involving months should encourage more comprehensive sampling of ice terrains. Furthermore, if detailed ice studies are impracticable in the field because of insufficient time, lack of necessary equipment, or lack of specialized personnel, specimens can be returned home for study.

INTRODUCTION

The difficulties involved in the preservation of ice specimens are partly responsible for the slow progress in the study of the physical characteristics of this interesting rock. These difficulties are particularly serious in places like Antarctica, where field parties are frequently away from base for weeks or months at a time and the problems of space and weight vie in importance with that of preservation of specimens. Yet, until methods are devised whereby small ice samples can be transported over large distances with nearly the ease with which other rocks are handled, the variety of specimens needed for comprehensive ice studies will be lacking. Only when the problems of transportation have been solved will specialists in well-equipped laboratories at home have access to samples from diverse and distant environments. As it is, expedition geologists are often compelled by circumstances to embark on ice studies in spite of the fact that their principal interests lie in other directions.

As a member of the recent Navy Antarctic Expedition, 1946–1947, under the technical command of Rear Admiral Richard E. Byrd, the writer was able to conduct a few simple experiments bearing on this problem. The experiments were designed to indicate the minimum size of ice specimen that needs to be collected in order to insure safe transport back to the United States. Unfortunately, there were no long-range field trips during the present expedition, the entire study being conducted at Little America and in the ship's refrigerator en route home. The tossing of the ship during the stormy return passage, however, subjected the specimens to gyrations probably fully as bad as those that are encountered on sledges or tractors in the field.

The brief period of preparation available to the writer before the departure of the expedition, as well as the need for consideration of other problems, allowed little time for contemplation of methods and acquisition of equipment for this study. Methods were devised en route, and makeshift equipment was salvaged or built from odds and ends aboard ship.

[1] Published by permission of the director, United States Geological Survey.

Hence no pretense at thoroughness is made. Recommendations are presented as guides for further study.

Most of the specimens herein described are now in storage in a refrigerating plant in Washington, D.C. They will be examined periodically for further wasting.

EQUIPMENT

Cylindrical tin cans of assorted sizes, with large circular lids, were selected for airtight storage (pl. 1, *A*). All paint was removed from the cans to permit maximum reflection of incident radiation. The large lids permitted the storage of specimens nearly as wide as the diameter of the cans.

A framework was built into one of the larger cans to house a specially built case for thin sections (pl. 2, *A*). This case was capable of holding, in horizontal position, ten 2×3-inch glass slides at intervals of $\frac{1}{2}$ inch. Two small blocks of ice cemented to glass slides were also stored in this container. Plate 2, *A*, also shows a second slide box, not airtight, which provided a basis for comparison of results with the airtight container. All containers were stored in a large wooden case, and the interstices were packed with snow. This wooden box was sheltered under an outdoor work bench at Little America (pl. 1, *A*) and later was stored in the ship's refrigerator.

During the stay on the ice, all thin sections were stored in the metal, airtight container. The container was filled with kerosene, which has proved to be an effective deterrent to evaporation (Demorest, 1940, p. 27). When the container was transferred to the ship, however, it became necessary to discard the kerosene for fear of imparting an odor to the foodstuffs in the refrigerated room.

TEMPERATURE CONDITIONS OF STORAGE

The maximum temperature recorded on the ice during our stay was 28° F. Thus, at no time were the specimens subject to melting temperatures. Other expeditions, however, have recorded temperatures above the melting-point at Little America. Although these were rare occasions of short duration, collection of specimens of somewhat larger size than would otherwise be collected might be advisable. The melting could be kept to a minimum or eliminated, however, by packing the containers in snow, which serves as a good insulator.

On shipboard the specimens were kept in a refrigerated storage room at a temperature of about 16° F. The ice studies which began at the base camp were continued in this room, something that could not have been done if the specimens had been stored in a small deep-freeze unit and a refrigerated room had been unavailable.

THE SAMPLES

Samples used in the study included specimens of the névé of the Ross Shelf Ice, ice from crevasses in the pressure folds of the Bay of Whales, and delicate sublimation crystals from the interior of the buildings of Little America II and III.

The density of the névé samples ranged from a little over 0.4 to about 0.54. Pore spaces were large, numerous, and frequently connected, allowing relatively easy penetration of air to a depth of probably 1–2 mm. One, and possibly both, of the crevasse specimens had a density of 0.87. Pore spaces were smaller, fewer, and more isolated than in the névé; hence these specimens exposed less surface to evaporation. No density determinations of the ice crystals were made.

PLATE 1

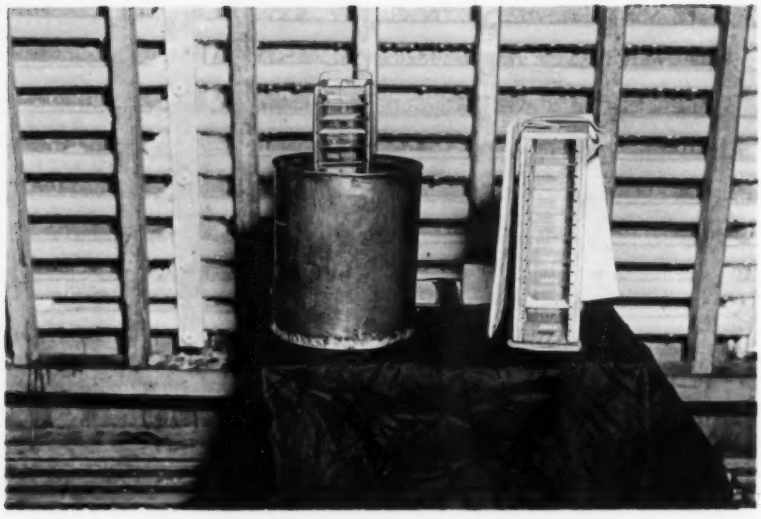
A

A, Work bench at Little America IV. Note the type of can used for airtight storage of specimens. The snow packing was used for all specimens except those mounted on glass slides. The double-bladed hacksaw, with blades 1½ inches apart, was designed by the writer to speed the preparation of thin sections 1½ inches square.

B

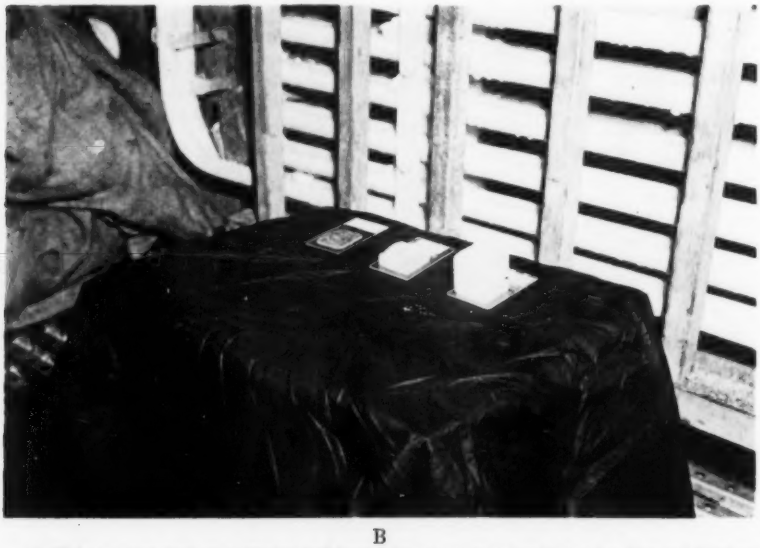
B, One of four sublimation crystals brought back from Little America. Length along diagonal, nearly 5 inches. This specimen, which showed no discernible evidence of wasting when examined in Washington, unfortunately broke while being prepared for rephotographing. The several fragments, as well as three other crystals, are in storage in Washington.

PLATE 2



A

A, The metal can served for airtight storage of thin sections. A housing within the can holds the slide-box snugly and prevents it from moving about. When in place, the slide-box rests entirely within the can. The box on the right is not airtight. Preservation results were compared with those obtained with the airtight can.



B

B, Thin section and two blocks fashioned from pressure ice from the Bay of Whales in late January, 1947. Photographed on arrival in Washington, April 14, 1947. The specimens suffered practically no deterioration in transport.

Twelve thin sections, 1½ inches square and 0.5–1.0 mm. thick, which had been prepared for petrographic study, were used in the preservation tests. Four of these were prepared at the base camp in the last few days of January. The remainder were made in the ship's refrigerator between February 10 and February 18. The thin sections were divided into two groups, the first of which was stored in the airtight can, the other in the case which was not airtight. A 1½-inch cube of crevasse ice and another piece 1½ inches square and approximately ½ inch thick (pl. 2, *B*), were also stored in the airtight container. The thin sections of ice, as well as the larger pieces, were cemented to the glass slides by application of water under freezing conditions. An orientation symbol was scratched on the upper surface of both the 1½-inch cube and the thinner fragment. The scratches were about $\frac{1}{16}$ inch deep.

In addition, irregular specimens of névé and crevasse ice were stored in individual airtight cans. Symbols were scratched on one or more faces of each specimen so that a rough estimate of the amount of wasting could be made.

The ice crystals (pl. 1, *B*), as well as all other specimens except the thin sections, were packed in snow in their containers, both to provide insulation and to absorb shocks encountered in transport. On arrival in Washington, the snow was easily brushed from the specimens.

RESULTS

Of the 12 thin sections that had been prepared up to February 18, those that had been stored in the airtight container fared best. Of these, the sections prepared from the denser, less porous, and less pervious pressure-folded bay ice suffered the least wasting, showing only rounding of the corners. The 1½-inch cube and the ½-inch thick section were practically intact on arrival in Washington on April 14 (pl. 2, *B*). Such wasting as did occur in the thin sections and glass-mounted ice fragments probably took place on the numerous occasions when the specimens were exposed for study in the ship's refrigerator. This is suggested by the lack of wasting of the delicate ice crystals which were not exposed during the return trip. However, the ice crystals were packed in snow, the influence of which on the rate of wasting was not determined.

The sections stored in the slide box which was not airtight wasted rapidly. By March 24, 5–6 weeks after their preparation, only one was still a 1½-inch square, although the corners were rounded and it had thinned appreciably. The remainder had wasted to remnants down to about ½ inch in diameter. The well-preserved section, which may have been a little thicker than the others to start with, was transferred at this time to the airtight container. On arrival in Washington, 3 weeks later, its margins had contracted only about ⅛ inch, in spite of the thinness of the section.

It is clear that preservation of thin sections of ice and transport over large distances are perfectly feasible if airtight containers are used and the storage temperatures are below freezing. Although the temperature of the ship's refrigerator was maintained at about 16° F., higher temperatures would probably not have appreciably affected the results of the tests.

A large part of the evaporation that did take place when the containers were open could probably have been eliminated by covering the thin sections of ice with cover glasses and sealing the edges

with vaseline, a procedure used by Perutz and Seligman (1939, p. 339). Unfortunately, the writer did not use such protected thin sections in his preservation tests. Warner (1945, p. 173) found that evaporation took place whether or not a covering was used, but he does not mention the steps taken to insure exclusion of air.

Although kerosene is an effective deterrent to evaporation, it has the disadvantage that, if frozen water is used as a cement and there are any chinks in the cement, the kerosene will be drawn into the section by capillarity. The large liquid bubbles beneath the section are particularly disturbing in petrographic examination. Eventually, the cement becomes weakened sufficiently so that the section comes loose. This situation is apparently aggravated by agitation of the kerosene during transport.

All irregularly shaped specimens were packed in snow in airtight cans. As pointed out earlier, markings were scratched to a depth of about $\frac{1}{16}$ inch on one or more surfaces so that a rough estimate of wasting could be made. In all cases, wasting was negligible, the markings having changed little if any on arrival in Washington, $2\frac{1}{2}$ months later. However, there was slight rounding of edges. Even this might have been prevented by application of vaseline or by soaking the snow in the containers with kerosene to eliminate air space. The use of vaseline or kerosene might also forestall possible sublimation growths.

Four delicate ice crystals (pl. 1, *B*) were also packed in snow in airtight containers. These crystals revealed no discernible evidence of wasting on examination in Washington. Unfortunately, in manipulating the crystal shown in plate 1, *B* for rephotographing, it broke into several pieces. Superficially, however, it seemed no different in Washington on April 14 than it did at Little America on January 28.

SUMMARY AND CONCLUSIONS

Small ice specimens, including thin sections and delicate ice crystals, were brought back intact from the Antarctic, a distance of about 10,000 miles by the route traveled. On arrival in Washington, the specimens had been in storage for periods ranging from 2 to $2\frac{1}{2}$ months. Such long-distance transport with little or no wasting requires packing in airtight containers. Thin-section trays should be of the type in which the slides rest horizontally. The trays must rest in a framework which prevents play. Jostling of hand specimens of ice is easily prevented by packing in snow, which also serves as a good insulating medium. Storage in temperatures well below freezing is advisable, inasmuch as evaporation is proportional to the temperature. Unfortunately, experiments involving controlled temperatures were not possible on this expedition; hence no suggestions on optimum storage temperatures can be made. It is interesting to note, however, that well-stored and sheltered specimens showed no ill effects at Little America when the temperature rose to 28° F. The results obtained by Perutz and Seligman (1939) recommend the use of vaseline as a deterrent to evaporation. Kerosene has certain disadvantages, unless precautions are taken.

For ordinary petrographic studies, ice specimens no larger than hand specimens need be collected even from such distant places as Antarctica. Specimens of this size can be collected in relative quantity without exceeding the space or weight limitations under which Antarctic field

parties operate. Thus opportunity is afforded to sample the ice of the Antarctic with nearly the same degree of thoroughness as that with which bedrock is sampled.

The feasibility of transporting small specimens of ice over great distances opens up interesting possibilities. Heretofore, ice studies had had to be completed before expeditions left the Antarctic. This sometimes led to haste in completing studies. Now part or all of the studies involving loose samples of ice can be postponed until arrival at home. This has the added advantage that, if the collector is not himself qualified to undertake such studies, he can submit his specimens to specialists. In any event, research would benefit from the superior equipment and the better working conditions found in the laboratories at home.

If studies other than petrographic are contemplated, larger specimens of ice would probably be needed. This would reduce the number of specimens which field parties could collect. However, this limitation applies as well to rocks other than ice.

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STUDIES FOR STUDENTS

SECONDARY TILT: A REVIEW AND A NEW SOLUTION

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INTRODUCTION

The problem of secondary tilt is one familiar to most geologists who have worked in the field. The primary dip of cross-laminations may be intensified or dissipated as a result of tilting of the strata of which they are a part; strata underlying many angular unconformities are secondarily disturbed by postunconformity movement. Often it is desirable to know, in the field, the attitude of these and other primary and secondary structures prior to later disturbances to which they may have been subjected.

In this paper methods of solution of the "two-tilt" problem that have come to the writer's attention are very briefly reviewed. The new method of solution presented was devised in the course of field study involving cross-laminated beds.

REVIEW OF SOLUTIONS

The earliest published solutions known to the writer are those of Harker (1884, pp. 154–162), who presented both graphic and algebraic methods. In the former, tangents are scaled as vector quantities, making visualization of the problem difficult.

Mertie (1922, pp. 572–574) referred the problem of secondary tilt to spherical trigonometry and showed how it may be solved by use of formulas based on spherical triangles. Fisher (1937, pp. 340–351) called attention to, and represented, Harker's original method of solution and later showed how the problem of two tilts may be solved by use of the stereographic projection (1938, pp. 1261–1271). Rapid preparation and interpretation of stereographic projections are made possible by use of a projection protractor devised by Fisher (1941, pp. 292–323, 419–442).

Bucher (1944, pp. 191–212) later presented the stereographic-net method of solution; in an appendix to the same paper, Fisher suggested a simplification of Bucher's method.

Spieker (1938, pp. 1255–1260) demonstrated that Harker's method is reliable only within certain narrow limits and that the illustration chosen by Fisher (1937) happened to fall within those limits. Spieker corrected Harker's original solution to make it applicable to all situations involving secondary tilt and presented another solution in which cotangents are plotted rather than tangents.

Corbett (1919, pp. 610–618) suggested a method by which contour maps may be used to work out original local structure in beds underlying an unconformity. By calculations involving the general (regional) strike and dip of rocks above and below the unconformity and the local structure of those above (with maps of both), the subsurface elevations of points on the strata as they were before secondary tilting may be obtained. With these as datum points, the original structure may be contoured. The method obviously has possibilities as applied to subsur-

face work but is somewhat unwieldy and impracticable for many stratigraphic and structural applications.

The various methods of solution referred to have certain disadvantages in that all require the use of data which may not always be available or make use of complicated formulas or involve map construction. However, for solution of problems involving numerous secondarily tilted units, such as the structural analyses of cross-lamination, the stereographic net is probably the best tool yet presented.

SUGGESTED NEW METHOD OF SOLUTION

The solution here presented is completely graphic and makes use of only simple geometric principles. No materials are needed that are not normally a part of every field geologist's equipment. Accuracy is dependent only on careful work and the scale used (fig. 1).

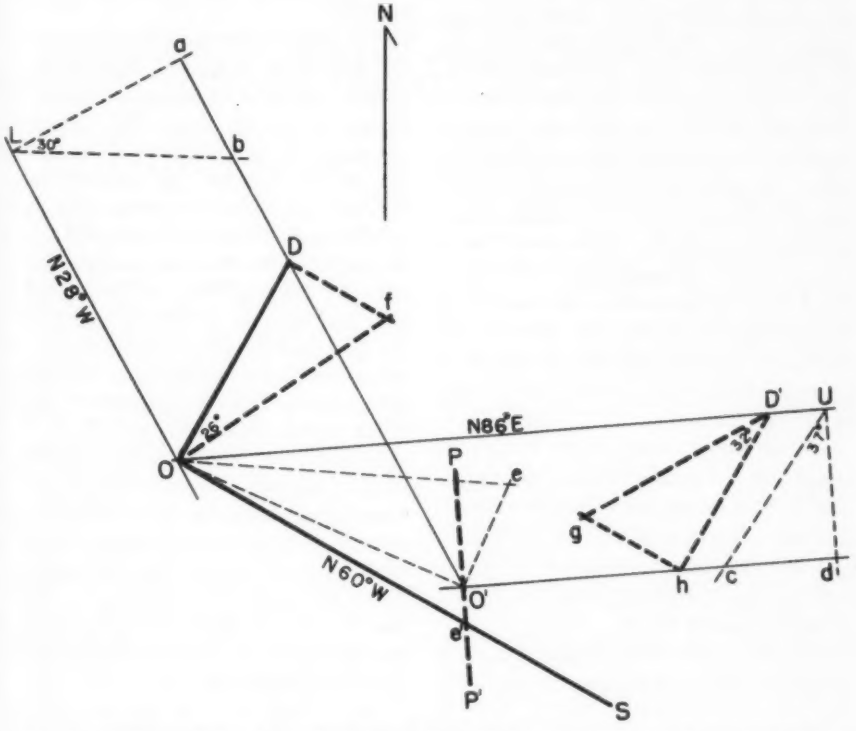

FIG. 1.—Graphic solution of a secondary-tilt problem. The fine-line construction is that used in determining the line of intersection of the two dipping planes. The construction in heavy lines is that by which the dip of *LOO'a*, prior to tilting of *UOO'd*, is determined.

Problem.—Strata striking N. 28° W. and dipping 30° to the northeast are unconformably overlain by strata which strike N. 86° E. and dip 37° to the south. Determine the strike and dip of the un-

derlying strata prior to the deformation that tilted the overlying beds.

Solution.—With the method suggested by Reid (1909, p.174) and modified by Lahee (1941, p. 684), the strikes of the strata beneath and above the unconformity are represented by lines *LO* and *OU*, respectively; the angles of dip are represented by *aLb* and *cUd*. The difference in elevation of the lower beds between *L* and *a* (arbitrarily chosen) is *ab; cd* equals *ab* and is the difference in elevation of the upper beds between *U* and *d*. Thus *aO'* and *O'd* are the strike lines on a plane at *ab* (or *cd*) distance vertically below the plane of *LO* and *OU; LOO'a* and *UOO'd* are, therefore, planes representing the attitudes of the strata below and above the unconformity. Line *OO'* is the horizontal projection of the line of intersection of these two planes; *O'e* equals *ab* (and *cd*); and *O'Oe* is the angle of inclination of the line of intersection of the two planes.

If the entire construction is rotated about line *UO* until *UOO'd* is horizontal, line *Oe* (the actual line of intersection of the two planes) will also be horizontal and will coincide in direction with the strike of the underlying beds before they were secondarily tilted.

Line *PP'* is the trace of a vertical plane normal to *UO* and passing through *O'*. Before rotation, *e* is vertically below *O'* and in the plane *PP'*. Line *Oe'* equals *Oe* and occupies the new position of *Oe* after rotation; and *O'e'* is the projection in plan view of the arc through which *e* moves as rotation progresses. Thus *Oe'*, or *OS*, is the strike of the strata beneath the unconformity prior to the second tilting; and the dip direction is along line *OD*.

The original dip along *OD* is found by the use of the apparent dips of planes *LOO'a* and *UOO'd* along *OD*. Line *Df* equals *ab*, and the angle *DOf* is the present angle of apparent dip, along *OD*, of the strata beneath the unconformity. Similarly, the present apparent dip of the overlying strata in direction *DO* is the angle *gD'h*. As rotation progresses, the value of angle *gD'h* is added to that of *DOf*, and the sum of the two angles is the dip of the lower strata along *OD* prior to the second deformation. Thus the underlying strata (*LOO'a*) dipped 58° to the north with a strike of N. 60° W.

In problems in which the dip directions of the two sets of strata are in the same hemisphere, the dip of the older set prior to the last deformation is equal to the *difference between* the two apparent dips. Normally, the direction of dip—that is, whether from *O* to *D* or from *D* to *O* in figure 1—may be readily determined by inspection.

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GEOLOGICAL NOTES

A NEW SPECIES OF EMBOLOMEROUS AMPHIBIAN FROM THE PERMIAN OF OKLAHOMA

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ABSTRACT

Archeria victori, sp. nov., is an amphibian from the Wellington (Lower Permian) collected near Eddy, Oklahoma. That the specimen represents an animal new to science is borne out by the character of the teeth and the absence of small secondary teeth on the lingual wall of the jaw.

INTRODUCTION

The specimen described in this paper (U.O.M. 2-1-S8), was collected in the McCann rock quarry in the Wellington sandstone, 2 miles northeast of Eddy, Oklahoma. It consists of the posterior portion of the right mandible. Associated with it were numerous excellent skulls, and one nearly complete skeleton of *Captorhinus*, a few *Dimetrodon* bones, one *Trimerorhachis* skull, an *Eryops* left mandible, abundant abdominal scales of Archeria, and some indeterminate remains. All these forms are typically Permian and doubtless occur at a much later date than the *Cricotus* material described by Cope (1875–1876, pp. 404–411) from the Pennsylvanian of the Danville, Illinois, area. It seems likely that the Oklahoma specimen more nearly corresponds with the *Cricotus* material described by Cope (1878, pp. 505–530) from the Permian of northern Texas. Dr. A. S. Romer (1947, p. 269) has recently assigned the Texas cricotid forms to the genus *Archeria* on the ground that they are all from about the same horizon, that they are geologically much younger than the Illinois forms, and that "no adequate morphological distinctions are known."

The occurrence of all the forms mentioned above in the Briar Creek bone-bed in Archer County, Texas, suggests corresponding age. At that location Case (1915, p. 157) reported the following forms in association with "*Cricotus*":

Dimetrodon incisivus Cope
Dimetrodon sp., small form
Theropleura sp.
Clepsydrops natalis Cope
Diadectes sp.
Diadectes maximus Case
Bolosaurus striatus Cope
Poliosaurus sp.
Archeria robinsoni genus et sp. nov.
Eryops megacephalus Cope
Cricotus heretoclitus Cope
Diplocaulus sp.
Trimerorhachis (?) sp.
Dorsal plate of a member of *Dissorhophidae*
Dermal scutes of an amphibian
Numerous bones of amphibians and reptiles, many belonging to new forms, but some undoubtedly belonging to known forms described from incomplete material

Van Vleet (1901, p. 60) lists the following forms at Orlando:

Diplocaulus magnicornis Cope
Diadectidae, Gen. indt.
Pariotichus incisivorus (?) Cope
Labyrinthodont
Trimerorhachis

And, in 1927, Smith (1927, p. 37) reported the following forms from the same locality:

Pleuracanthus cf. compressus
Gnathorhiza pusilla
Diplocaulus salamandroides
Trimerorhachis sp.
Crossotelos annulatus

Romer (1935, p. 1625) reports the presence of *Archeria* in each formation of the Wichita group and in the Pueblo formation of the Cisco group in the Texas region. The genus has also been reported at Morrison (Smith, 1927, p. 42) (Wichita), Pond Creek (Smith, 1927, p. 4) (Clear Fork), and Orlando (Case, 1901, p. 63) (Clear Fork), Oklahoma, under the name of *Cricotus*. The writer has collected fragmentary remains of the genus in Wichita beds in Cotton, Stevens, and Jefferson counties, Oklahoma, where the formations are approximately the

equivalent of the Texas beds referred to by Romer. The Eddy find appears to extend the vertical range of this genus into the Clear Fork, although the position of the bone-bearing beds is 4 feet and 4 inches below the top of the Wellington, as shown by Clark and Cooper (1927, p. 10). Numbers 11 and 12 in the section given in table 1 are considered an outlier of the Garber.

The measured section described in table 1 is generalized, but the Wellington portion was taken on the side of the quarry from which the bones came.

TABLE 1*

MEASURED SECTION IN ROCK QUARRY IN NW¼ SEC. 16, T. 26 N., R. 2 W., KAY COUNTY, OKLAHOMA

MATERIAL	THICKNESS	
	Feet	Inches
13. Top soil	2	0
Garber sandstone:		
12. Sandstone, red, bedded, very fine-grained, ripple-marked; weathers into thin flagstones	1	6
11. Sandstone, dark red, tightly cemented, hard		6
Wellington formation:		
10. Clay-shale, gray, red and yellow banded	1	2
9. Limestone, gray, very finely crystalline; grades locally into gray shale		0–4
8. Shale, dark gray, fissile	1	4
7. Clay-shale, drab	1	6
6. Sandstone, red with gray streaks, hard, with abundant small bone fragments; grades northward into gray silty shale (fossil horizon)		6
5. Sandstone, red, soft; grades northward into gray shale		6
4. Clay-shale, gray-green, soft	3	0
3. Limestone, gray, very finely crystalline		0–1
2. Clay-shale, red	4	0
1. Limestone, light gray, finely crystalline, with abundant vugs lined with calcite crystals		4
	16	6

* Measured by Ralph Disney and Lynn Jacobsen at the writer's request. This section is in the interfingered "zone of color change," where within a short distance the Wellington formation changes facies laterally southward from gray shale and mudstone with minor amounts of limestone to red shale and sandstone.

DESCRIPTION OF MATERIAL

It is regrettable that the specimen is not complete. Nevertheless, the find is beautifully preserved, and it seems advisable to present descriptions at the present time with the hope that it will add to the knowledge of this poorly known form. Although only about two-thirds of the posterior portion of the right mandible was secured, the articular is complete and perfect, and the sutures are distinct. The complete length is approximately 345 mm., about the same size as the largest individuals of this genus discovered. The posterior half of the jaw (this specimen) of *Archeria* resembles both the pelycosaurs and the cotylosaurs, but, on the basis of the teeth, it is clearly of embolomerous affinities.

The height-width ratio of the specimen is distinctive, since it is relatively high and thin. The sides are nearly parallel. This specimen is, however, more obtuse in the angular region than any known species of this genus. The sutures are, in general, distinct, and the bones agree in shape only with the high-jawed amphibians.

In this specimen the articular, the prearticular, and the two coronoid bones are complete, while only a portion of the dentary, surangular, angular, and splenial are present. As in "*C.*" *heteroclitus*, there are two Meckelian fossae. The bones surrounding the large Meckelian fossa are distinct and complete, so that there is no doubt as to the size of the opening. The anterior end of the infra-Meckelian fossa can be only approximated because of the termination of the specimen at about that place. This opening is, however, long and narrow. Just anterior to and slightly above the infra-Meckelian fossa there appears to be the posterior end of a nutrient foramen. If this is the case, the present fossil corresponds in that particular to "*C.*" *heteroclitus*.

Posteriorly, the jaw is deep and quite obtuse—a feature that distinguishes it from all known species. The articular is complete and stands well above the tooth row.

The dentary of this specimen covers about one-fourth of the exterior surface of the jaw and is bounded below by the surangular. Posteriorly, it narrows toward the top of the jaw and is bounded above by the coronoid. As in the pelycosaurs and most amphibians, the thick upper surface of the dentary serves as attachment for

the lateral tooth row, which is bounded internally by the coronoid bones in this specimen.

The tooth row is not complete; but sixteen teeth are present, with gaps for three others and a stub of still another at the anterior end. The total number of teeth, therefore, cannot be determined; but the general form of the jaw suggests that there were between 45 and 52. There is no evidence of small teeth on the coronoid as in "*C.*" *heteroclitus*. The teeth preserved in this specimen are not differentiated as in most amphibians and are closely spaced, vertical, and remarkably uniform in size. The wear on the cutting edges of the teeth suggests a herbivorous diet, although Case (1911, p. 148) has suggested crocodilian habits for the beast. The teeth are decidedly chisel-shaped, as described by Case (1915, p. 160), and, in this respect, they resemble the marginal teeth of *Captorhinus*.

The teeth are peculiar in that they are flattened ovoid in cross section at the base, the greater diameter being perpendicular to the tooth row. This character is less pronounced in an anterior direction. Laterally, there is not much difference in diameter at the base and in the upper portion of the teeth. On the lingual

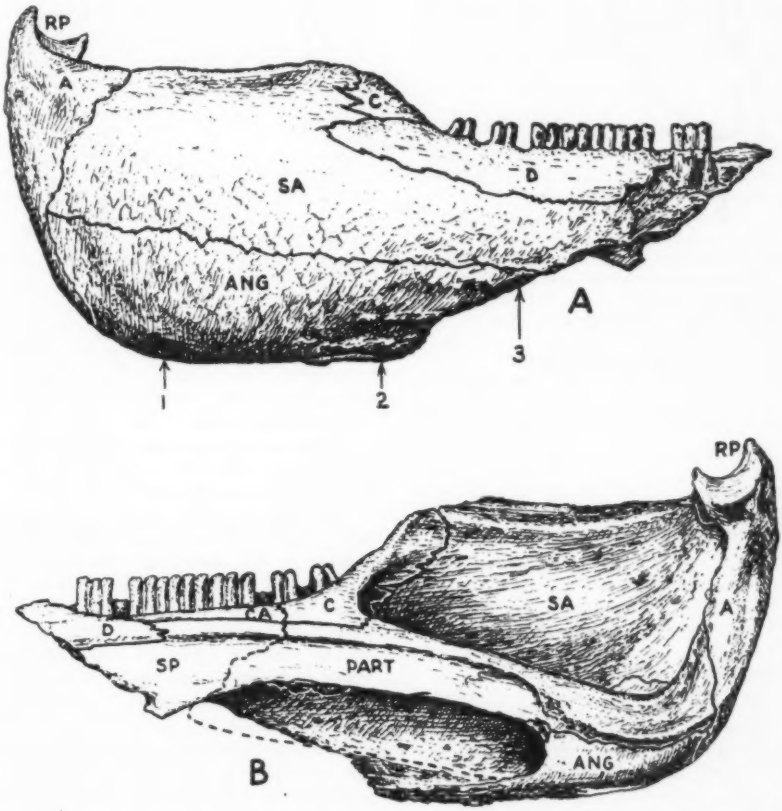

FIG. 1.—Two views of right mandible of *Archeria victori* sp. nov., from Eddy, Oklahoma. *A*, lateral view, length of fragment 192.2 mm., or about 7.5 inches. Figure is three-fifths natural size. *B*, internal view. *A*, articular; *ANG*, angular; *C*, coronoid; *CA*, anterior coronoid; *D*, dentary; *PRA*, prearticular; *RP*, retroarticular process; *SANC*, surangular; *SP*, splenial.

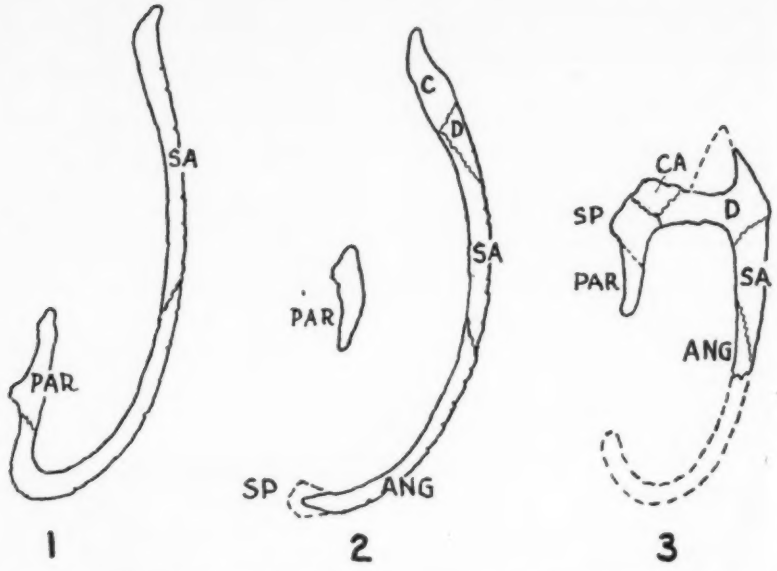

FIG. 2.—Sections through the jaw of *Archeria victori* sp. nov. at points *1*, *2*, *3*, indicated in *A* of Fig. 1. Each illustration natural size.

TABLE 2

MEASUREMENTS

	Width (Mm.)	Length (Mm.)
Angular	38.2	90*
Articular	28.2	49.5
Anterior coronoid	4.4	38.9
Coronoid	18.8	48.1
Dentary	15.3	95
Prearticular	16.6	100.4
Retroarticular process	30.2	.....
Surangular	48.4	141
Splenial	26.5	43.7

* Length of existing fragment.

side there are vertical striations that extend from the base to within 3 mm. of the top of the teeth. Similar striations occur on the teeth of some of the cotylosaurs and in all known species of "*Cricotus*."

When viewed from the side, the tops of the teeth form an almost straight line, and the length of the individual teeth ranges from 9.1 mm. at the posterior to 12 mm. at the anterior end. The width departs slightly from 3.6 mm. There is posterior overhang of the cutting edge on only a few of the teeth, a feature that is more pronounced in "*C.*" *heteroclitus* as described by Case (1915, p. 160).

There are two small coronoids, the true coronoid being visible above the posterior end of the dentary in an outside lateral view.

The angular forms the posteroventral portion of, and terminates posteriorly on, the outer surface of the ramus, continuing as a deep ventral keel on the posterior half of the jaw below the dentary and surangular. This bone is visible internally through the Meckelian fossa.

The articular forms the posterior margin of the ramus and is bounded externally by the surangular and angular and internally by the Meckelian fossa and the prearticular. The articular forms a margin or ventral keel between the inner and the outer surfaces of the jaw and extends upward, forming the retroarticular process, which in most other amphibians is well above the level of the tooth row.

The prearticular is a narrow bone, bounded posteriorly by the articular and angular and anteriorly by the coronoid and splenial. The pre-

articular also forms the lower boundary for the Meckelian fossa and serves as the upper margin of the infra-Meckelian fossa, forming the median wall of the posterior part of the Meckelian canal. This bone curves downward slightly and forms a bar across the inner surface of the jaw in a characteristic manner. Just anterior to the infra-Meckelian fossa, seen as a broad notch just below *SP* in *B* of figure 1, there appears to be an indication of a foramen in the splenial (see table 2).

CONCLUSION

That this specimen is *Archeria* is suggested by the following characteristics: (1) the presence of like bones and similarity of their arrangement in the mandible; (2) the thin, deep character of the posterior portion of the jaw; (3) the arrangement of the teeth in a groove along the dentary; (4) the fact that the teeth are not differentiated in this portion of the jaw; (5) the position of the retroarticular notch and its resemblance to that in known specimens; (6) the parallel arrangement of the cutting edge of the teeth; and (7) the chisel shape of the teeth.

Specific characteristics of *A. victori* are: (1) the absence of small teeth on the coronoid; (2) the closer spacing of the teeth; and (3) the extreme bluntness of the posterior end of the jaws.

The species name, *victori*, is in honor of Professor Victor E. Monnett, who, as head of the Department of Geology at the University of Oklahoma, has been instrumental in the advancement of scientific research at the University.

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AN EARLY REPORT OF ANCIENT LAKES IN THE BONNEVILLE BASIN

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Most histories of geology credit Howard Stansbury with the discovery, in 1849, of field evidence indicating ancient higher lake levels in the Bonneville Basin (Stansbury, 1852; Gilbert, 1890, pp. 12–13). Stansbury's published reports indicate that he not only recognized the evidence for the ancient lakes but reached a reasonable conclusion as to their magnitude.

Recent translation and annotation, by Herbert S. Auerbach, of the diaries of the Escalante party, which traversed eastern and central Utah in 1776, makes available evidence which very strongly suggests that Stansbury's discovery was anticipated by nearly three-quarters of a century.

In the diary of Silvestre Veléz de Escalante, "a Franciscan Friar of great intelligence and ability," under the date of October 2, 1776, is the notation ". . . . This place, which we named Llano Salado, because we found some thin white shells there, seems to have once had a much larger lake than the present one." Location of this site, according to Auerbach's reconstruction of Escalante's route (Auerbach, 1943, pp. 1–142) is between Sevier Lake and the Beaver Mountains, in Millard County, Utah. Esca-

lante's diary includes a description (October 3, 1776) of areas in which is exposed "white, dry, fine loam, which from far away looked like white linen spread out to dry . . . ." (Auerbach, 1943, pp. 75–76; map opp. p. 90).

Aerial and jeep reconnaissances of the general region show that Auerbach's site identification is in close accord with the data recorded by Escalante but that a second site, east of the Beaver Mountains and south of Clear Lake, not far from Neels Station on the Los Angeles and Salt Lake Railroad, fits the description equally well. At both sites, "white, dry, fine loam . . . ." (Gilbert's "white marl"), containing large numbers of shells (largely *Lymnaea bonnevillensis*), is present.

In view of the foregoing, it appears that Escalante discovered and reported evidence of high lake levels in the Bonneville Basin some seventy-three years before Stansbury. Present historical knowledge indicates that this is the earliest such report, as the Escalante expedition was the first official party to enter the Utah Desert.

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THE COMPOUND $CaO \cdot 2Al_2O_3$

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In the original work on the system CaO–Al_2O_3 (Shepherd, Rankin, and Wright, 1909, pp. 293–333) a compound $3CaO \cdot 5Al_2O_3$ was identified as one of the calcium aluminates stable at liquidus temperatures. The composition was determined by microscopic examination of synthetically prepared mixtures, fused in an iridium furnace. Two modifications of this compound were described—a stable, uniaxial modification, with $\epsilon = 1.651$ and $\omega = 1.617$, and an "unstable" biaxial form, with $\alpha = 1.662$, $\beta = 1.671$, $\gamma = 1.674$, and a $2V$ of $35° \pm 5°$. Individual crystals of the unstable modification were not observed; they all showed evidence of alteration to the uniaxial form, which proceeded inward from the periphery of the grains.

In 1937 two papers were published within a few weeks of each other, one by Lagerqvist, Wallmark, and Westgren (1937, pp. 1–16) working in Sweden, the other by Tavasci (1937, pp. 717–719) in Italy, in which the conclusion was independently reached that the $3CaO \cdot 5Al_2O_3$ formulation was incorrect, the ratio of lime to alumina in reality being 1:2. The conclusions of Lagerqvist *et al.* were based on X-ray evidence in conjunction with density determinations, as well as powder photographs. It should be noted that the Weisenberg determinations showed the material to be monoclinic, with $a = 18.82$ A, $b = 8.84$ A, $c = 5.42$ A, and $\beta = 107°8$. No optical determinations were made. Tavasci arrived at the $CaO \cdot 2Al_2O_3$ composition on the basis of examination of etched polished specimens with the ore-microscope. He found that crystallized melts of the 3:5 ratio were inhomogeneous, whereas a homogeneous single crystalline phase developed from a melt of the composition $CaO \cdot 2Al_2O_3$. This compound melted incongruently at a temperature of about 1765° C. No data on optical properties are given in Tavasci's paper, but he states that they are the same as those assigned by Wright to the 3:5 compound.

The American literature has virtually ignored this 1937 work; the only references containing any mention of even the possibility of the existence of the 1:2 compound known to the writer are (*a*) the A.S.T.M. X-ray diffraction data cards and (*b*) Bulletin 107 of the National Research Council: *Data on Chemicals for Ceramic Use*. Both these sources are merely compilations of data, and in neither is the question of composition clarified; the reader cannot be certain as to which formulation is correct or whether both compounds exist. The 1942 printing of the phase diagram for the system CaO–Al_2O_3–SiO_2 (Sosman and Anderson) shows the

compound $3CaO \cdot 5Al_2O_3$ as the calcium-aluminate in question, as does the diagram just published in the compilation of phase diagrams by the American Ceramic Society (Hall and Insley, 1947). The fact that the American literature is not up to date on this subject led the writer, in a discussion of solid-solution relations in a recently published paper (1947, pp. 381–404), to mention "a hitherto unknown 'molecule,' of the composition $CaO \cdot 2Al_2O_3$."

Some experimental work was done to help clear up the problem and as a check on the data of Tavasci and of Lagerqvist, Wallmark, and Westgren. A mixture of CaO (as $CaCO_3$) and Al_2O_3 in the 1:2 ratio was carefully prepared; it was thoroughly mixed, sintered at approximately 1500° C., crushed, remixed, and resintered. Portions of the mass were then fused in an oxygen-gas flame. Microscopic examination confirmed Tavasci's observation—a single homogeneous phase had crystallized from the melt. It is interesting to note that virtually complete reaction had taken place during sintering at 1500° C. in a relatively few hours, as seen microscopically. The optical properties are indeed those described by Wright (1909); the crystals are uniaxial (+), although, as Wright observed, some show a slight tendency toward biaxial development. The indices of refraction are $\omega = 1.617 \pm 0.002$ and $\epsilon = 1.651 \pm 0.002$ (sodium vapor light). No trace of the "unstable phase" was observed, although no special effort was made to obtain it. X-ray powder pictures show the same pattern as that obtained by Lagerqvist and co-workers. Their technique apparently did not permit the observation of diffraction lines of low θ value, however, as several lines are present that they did not find; one strong line of $d = 4.44$ (indices 020) and a moderately weak line of $d = 6.17$ (indices 200) are worth mentioning. Slight discrepancies in spacings might indicate that some refinement of unit-cell dimensions is desirable. No crystals sufficiently good for a Weisenberg determination were obtained.

The 1909 diagram of Rankin and Shepherd indicated that the 3:5 compound melted incongruently. In 1915 a revised diagram by Rankin and Wright (p. 11) has the compound as a congruent one, possibly as a result of the location of the boundary curve in the ternary system. The 1:2 ratio of lime to alumina places the composition point outside its stability field, however, if the boundary curve is correctly placed, and thus, as pointed out by Tavasci, $CaO \cdot 2Al_2O_3$ melts incongruently.

No additional information has been obtained as to the existence of an unstable modification. The X-ray diffraction powder picture obtained by Lagerqvist, Wallmark, and Westgren indicates that they dealt with the same modification here described, in which case the compound is monoclinic. Observation of optical properties alone would lead one to the conclusion that $CaO \cdot 2Al_2O_3$ is hexagonal or tetragonal.

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COMMUNICATIONS TO EDITOR

TRANSLATIONS OF RUSSIAN GEOLOGIC LITERATURE

The Committee on Russian Literature of the Geological Society of America is compiling a list of English translations of Russian geologic papers and books covering all fields from geophysics to paleontology. Information regarding manuscript translations is desired. It should be sent to the chairman of the committee, Ronald K. DeFord, Box 1814, Midland, Texas, or to the secretary of the Geological Society. Such information will be published in the forthcoming volumes of the *Bibliography and Index of Geology Exclusive of North America* and the *Bibliography of Economic Geology.*

PUBLICATION OF THE GEOPHYSICAL ABSTRACTS

The United States Geological Survey has resumed publication of the *Geophysical Abstracts* after a four-year interval, during which they were issued by the United States Bureau of Mines.

The *Geophysical Abstracts* are published quarterly as an aid to those engaged in geophysical research and exploration. The bulletin covers world literature on geophysics contained in periodicals, books, and patents. It deals with exploration by gravitational, magnetic, seismic, electrical, radioactive, geothermal, and geochemical methods and with underlying geophysical theory and related subjects.

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REVIEWS

"Pollen Profile from a Texas Bog." By J. E. POTZGER and B. C. THARP. (*Ecology*, vol. 28.) Pp. 274–280, 1947.

The pollen profile is from the Patschke Bog, which is located 40 miles east of Austin in eastern Texas. The bog is remarkable for its great depth—22 feet—considering the low latitude, and for containing pollen grains of *Picea glauca* (=*canadensis*), *P. mariana*, and *Abies*. Samples were taken at 1-foot intervals. The significant vegetational changes and the climatic conclusions are briefly discussed below.

Foot-levels 22–18 (at bottom of bog).—Maxima of *P. glauca*, *P. mariana*, and *Abies* indicate a cool climate. Increase of predominantly medium-sized grass pollens from a minimum at the bottom to a maximum at foot-level 18 suggests a change from moist to dry climate. However, tall-grass prairie species, which have large pollen grains, did not predominate at these or at any other levels in the bog.

Foot-levels 17–16.—Maxima of medium-sized grass pollens and of *Quercus* and absence or scarcity of boreal conifers suggest a warm-dry climate.

Foot-levels 15–10.—Abundance of *Alnus*, moderate representation of *Castanea*, and minimum of medium-sized grass pollens suggest a warm-moist climate. The present western range limit of *Castanea* is about 100 miles to the east of the Patschke Bog, where the rainfall is several inches greater.

Foot-levels 9–6.—Maximum of *Castanea* and abundance of small-sized grass pollens (moist-habitat forms) suggest a continued warm-moist climate.

Foot-levels 5–1 (at top).—Maximum of medium-sized grass pollens, increase of *Quercus*, appearance of *Carya*, and decline and disappearance of *Castanea* indicate a warm-dry age.

The present vegetation of the region consists of the oak-hickory association. Tall-grass prairie begins 15 miles to the west. The climate is moist subhumid, and mesothermal (Thornthwaite, 1941, pl. 3).

The climatic history is thus briefly as follows:

Foot-Levels	
0 = top	
	Warm-dry
5·5	
	Warm-moist
15·5	
	Warm-dry
17·5	
	Cool, moist, growing dry
22	

Picea glauca and *P. mariana* are boreal trees whose modern southern range limit skirts the southern tip of Lake Huron and passes through central Wisconsin and the northwestern corner of Minnesota, following in this region the July isotherm of 70° F. (U.S. Dept. Agr., 1941). The pollens of these spruces in the Patschke Bog thus occur over 1,000 miles, or 13 degrees of latitude, south of the modern ranges of the trees. Cones, wood, and twigs of *P. glauca* in Pleistocene beds at Percy Bluff, 35 miles north of Baton Rouge in Louisiana, show that this spruce once really grew in these low latitudes (Brown, 1938, pp. 67, 86). Since the present-day July temperature at the Patschke Bog is 84° F., the July temperature in the region then was perhaps some 14° F. lower than now. A Pleistocene lowering of the July temperature of 13°–14° F. (7°.28–7°.8 C.) is also suggested by pollens of *Picea* and *Abies* in Florida (Davis, 1946); by pollens of a vegetation strikingly resembling the modern transition forest of northern Minnesota at the bottom of the Jerome Bog northwest of Wilmington in North Carolina (Buell, 1946, pp. 14–15); and by fossil marmots 4,000 feet above their modern level of range in New Mexico (Stearns, 1942, pp. 867–878).

The relative thinness of the cool bottom beds in the Patschke Bog suggests that they hardly extend beyond the last, the Mankato, glacial maximum, making the whole bog represent the last twenty-five thousand years.

The climatic history inferred by Potzger and Tharp resembles that concluded by Paul Sears

and E. S. Deevey for the "Lateglacial" and the "Postglacial" of the northeastern United States, except for the absence of a relatively cool and moist age, Sears and Deevey's "Subatlantic," during the past few thousand years. Of course, a correlation requires that Sears and Deevey's time scale be stretched from ten thousand to some thirty thousand years.

The climatic history of the bog shows less agreement with that of the region extending from central Texas into the state of Washington. In this vast area the climate became increasingly warm and dry until a few thousand years before Christ; and since about 2000 B.C. it has been predominantly moderately cool and moist but with dry (and warm?) periods. This seems to be the proper characterization of the last stage in Arizona and western Texas, for the recent erosion-recorded droughts can have been but brief interludes, to judge from the rapidity of the modern erosion in the region. One drought, recognized in northern (Hack, 1942, pp. 58, 68) and southeastern (Sayles and Antevs, 1941, pp. 43, 46) Arizona, in Trans-Pecos Texas (post-Calamity erosion) (Bryan and Albritton, 1943, pp. 486–487), and on the Texas High Plains ("renewed deflation") (Evans and Meade, 1945, pp. 499–503) occurred perhaps A.D. 1276–1299 (1273–1300), when narrow tree-rings were formed on the Colorado Plateau (Schulman, 1947, pp. 2–8). Another drought, recorded by erosion in southeastern Arizona and probably by an unconformity in the Calamity formation in Trans-Pecos Texas, seems to have taken place early in the Christian era. Consequently, the now moist and moist subhumid eastern Texas and the semiarid and arid country from western Texas to Washington seem to have had opposite temperature trends during the last four thousand years and opposite moisture trends almost throughout the Postpluvial.

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ERNST ANTEVS

Elements of Soil Conservation. By HUGH HAMMOND BENNETT. 1st ed. New York: McGraw-Hill Book Co., Inc., 1947. Pp. x+406; figs. 114. $3.20.

The broader problems of soil wastage and methods of soil conservation are here effectively presented in nontechnical terms. About one-third of the book is devoted to erosion processes and their effects and the remaining two-thirds to standard methods of soil conservation. Questions for discussion are given at the end of each chapter, and at the end of the book there is a list of films and film strips which can be used to supplement the text. The book is suitable as an elementary text for both student and adult groups. Teachers of beginning geology may find it useful for supplementary readings. It does not in any sense replace the same author's *Soil Conservation* as a comprehensive reference book.

L. H.